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Abstract

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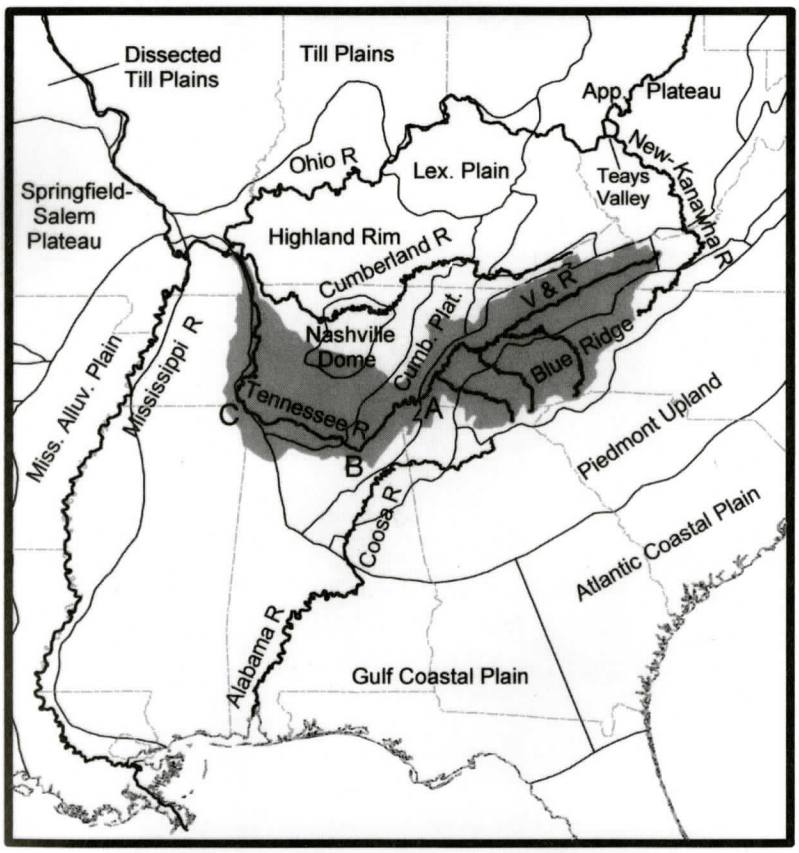

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DRAINAGE HISTORY OF THE TENNESSEE RIVER: REVIEW AND NEW METAMORPHIC QUARTZ GRAVEL LOCATIONS

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ABSTRACT

The strange course of the Tennessee River (TR) has mystified observers for more than a century. There are three seemingly unlikely course changes: 1) west of Chattanooga, Tennessee, where the river leaves the Valley and Ridge province and cuts through Walden Ridge; 2) near Guntersville, Alabama, where it leaves the southwestward trending Sequatchie anticlinal valley and assumes a northwesterly course; 3) near the juncture of the Alabama, Mississippi, and Tennessee borders, where it turns north to cross Tennessee and join the Ohio River. These abrupt changes suggest the possibility of former courses of the TR much different from the present one. Prominent in discussion of such courses is the "Appalachian River," a hypothesized river system that flowed through the Great Valley and continued its course southwestward to the Gulf Coast, through what is now the Alabama River system or via other routes farther west. An early argument for the course change 1) by stream capture of the Appalachian River by a westward-flowing tributary of the Sequatchie River during the late Tertiary appears weak; probably the Walden Ridge gorge is no younger than early Cenozoic, and may even a have been carved by a consequent stream of the Paleozoic Appalachians. To explain 2) and 3), several efforts have been made to show that these courses show adjustment to bedrock structure. Small Cenozoic crustal movements may have been equally as important in influencing the course, so such efforts cannot be conclusive. Evidence from deposits of former stream courses would be much more convincing. Heavy-mineral suites and metamorphic quartz (MQ) derived from the crystalline Appalachians are widespread on the coastal plain, attesting to an ultimate source in the Blue Ridge as early as Cretaceous. Although some efforts involving subsurface data have been made, most attempts to demonstrate former TR courses have been based on surface or near-surface gravel deposits. Such courses are likely to be no older than Miocene, as older surficial deposits probably would have been removed by erosion. We here give locations of more than 100 gravel sites in Alabama, Mississippi, and western Tennessee with high MQ contents. These locations provide constraints on possible former courses of the TR. A major problem is whether the MQ in the deposits came directly from the Blue Ridge via an ancestral TR or whether it came from the reworking of older formations containing MQ, particularly the Pennsylvanian sandstones and the Tuscaloosa Formation, by local streams. In some cases, this problem can be avoided by confining attention to MQ clasts that are cobble-size or larger, and/or to areas in which the Tuscaloosa lacks significant MQ. We suggest that future study of possible old TR courses begin with study of high terraces along the present TR, concentrating on dating and lithology studies of the deposits.

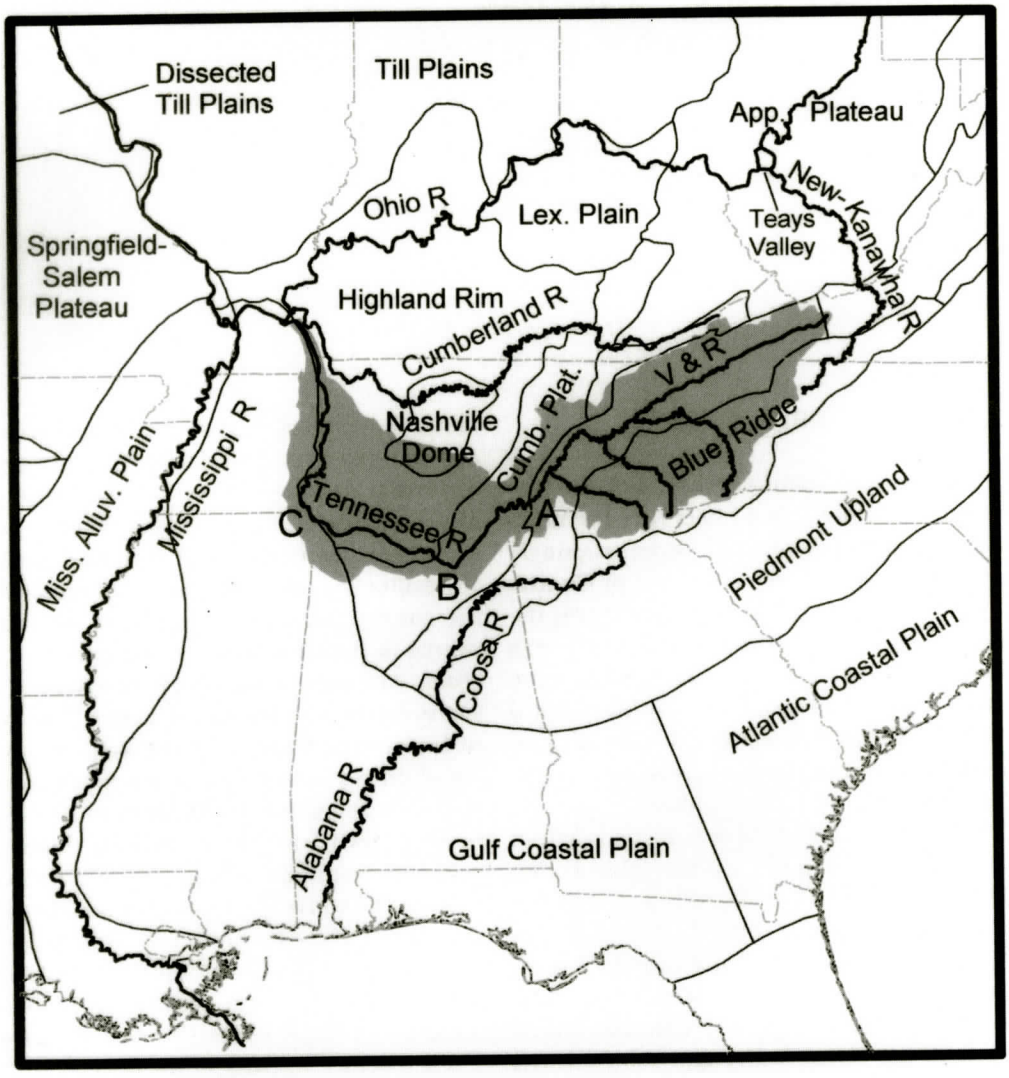

Figure 1. Regional setting of the Tennessee River, showing major rivers and physiographic provinces. Shaded area is drainage basin of Tennessee river.

INTRODUCTION

The course of the Tennessee River (TR) seems a geological anomaly. Arising in Virginia, the TR follows strike valleys in the Valley and Ridge province as far southwest as Chattanooga (Figs. 1-3). At this point, a continuation down the Great Valley to the coastal plain and thence to the Gulf of Mexico via what is now the Coosa-Alabama River system seems the path of least resistance. Instead, the river course abruptly turns west and cuts through the 300-m-high, sandstone-capped Walden Ridge (A in Fig. 1), entering the valley of the dissected Sequatchie anticline, although a divide only 75 m high separates it from headwaters of the Coosa-Alabama drainage system (Fig. 2). In the valley of the Sequatchie anticline, the TR again heads for the Gulf. However, when it reaches the vicinity of Guntersville, Alabama, instead of continuing a course into the headwaters of what is now the Black Warrior River, it once again takes

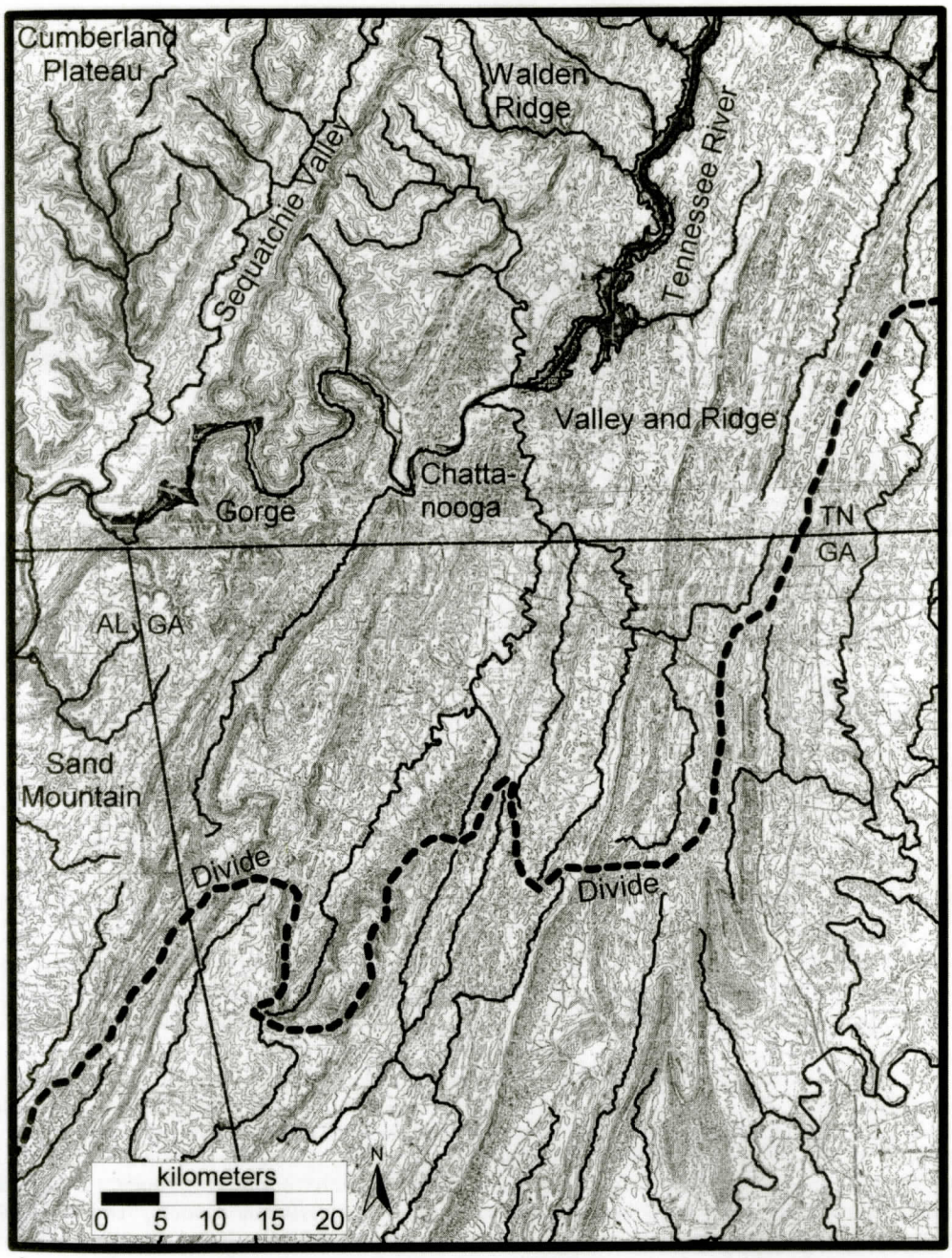

Figure 2. Drainage and topography in the vicinity of Chattanooga, Tennessee. Divide shown is between the Tennessee River system to the north and the Coosa River system to the south, and is only 75 m high in some places.

an abrupt right-angle turn to the northwest (B in Fig. 1), where it follows the strike of Mississippian strata and then assumes a more westerly course. At this point, the river course seems to be headed to the Mississippi River. Instead of continuing this westerly course, the TR takes its strangest turn of all. Near the Alabama-Mississippi border (C in Fig. 1), it turns north across

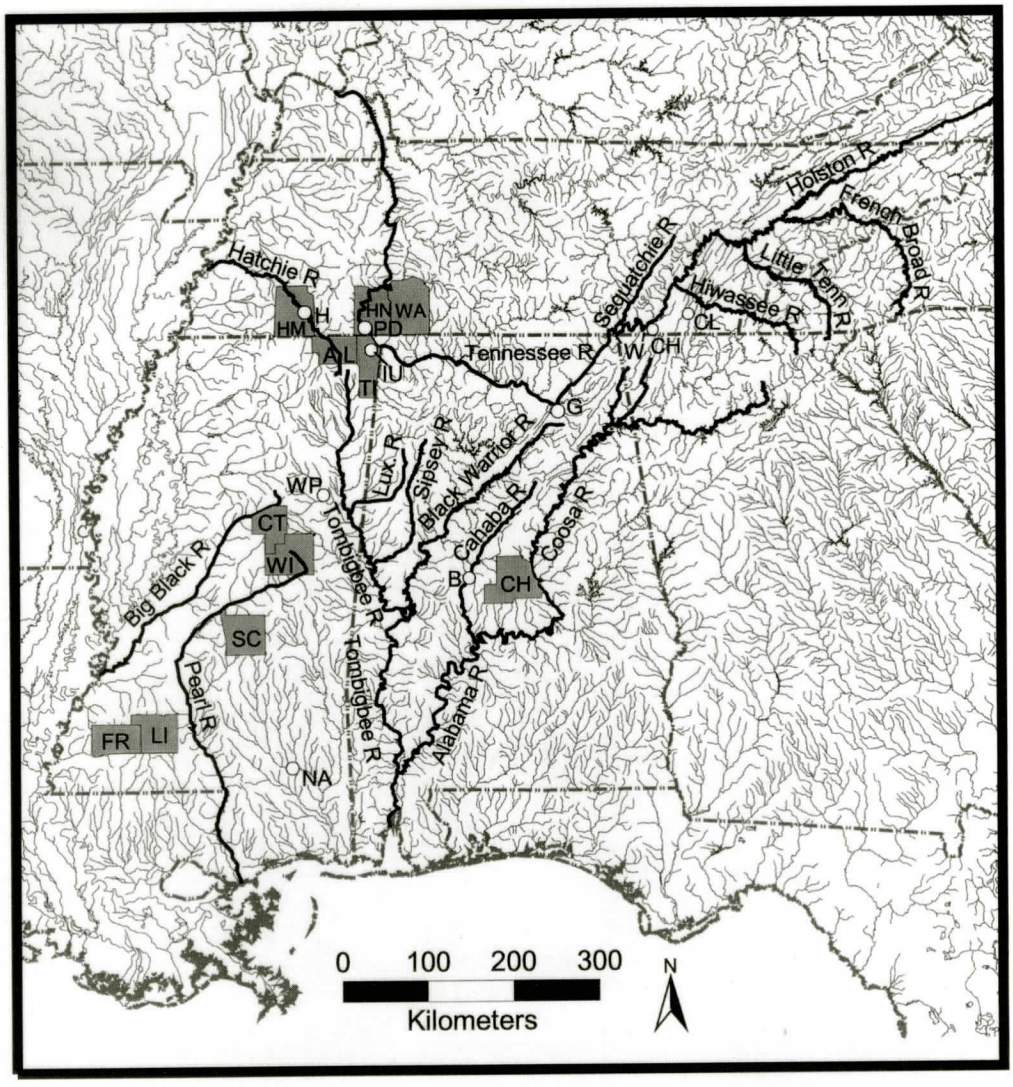

Figure 3. Location map showing rivers and sites discussed in text. Abbreviations: B = Brent, Alabama; Tennessee; CH = Chattanooga, Tennessee; CL = Cleveland, Tennessee; G = Guntersville, Alabama; H = Hebron, Tennessee; IU = Iuka, Mississippi; NA = New Augusta, Mississippi; PD = Pickwick Dam, Tennessee; WP = West Point, Mississippi; Lux. R = Luxapalila River, Mississippi. Shaded counties: AL = Alcorn County, Mississippi; CH = Chilton County, Alabama; CT = Choctaw County, Mississippi; FR = Franklin County, Mississippi; HM = Hardeman County, Tennessee; HN = Hardin County, Tennessee; LI = Lincoln County, Mississippi; SC = Scott County, Mississippi; TI = Tishomingo County, Mississippi; WA = Wayne County, Tennessee; WI = Winston County, Mississippi. Note that the Tennessee River formally begins at the junction of the Holston and the French Broad Rivers.

Tennessee and Kentucky until it finally joins the Ohio River.

Despite the striking nature of this geomorphic problem, after the early 1900's most papers addressing the course of the TR have done so only in passing; there have been very few studies devoted exclusively to it. This paper attempts to bring together the scattered previous

work on the TR, provide new data bearing on the problem, and suggest avenues of future work.

INITIAL WORK: THE APPALACHIAN RIVER AND WALDEN RIDGE

A former course for the TR from Chattanooga to the Gulf of Mexico via the Coosa River system (Fig. 1) was suggested as early as 1875 (Long, 1875, p. 16). The first geomorphological work on this question was by Hayes and Campbell (1894). Based on the depositional history of the coastal plain, they hypothesized that there had been two major episodes of crustal uplift and quiescence. Uplift preceding and during the earlier part of the Cretaceous was followed by a long period of baseleveling in the Late Cretaceous, during which a well-developed peneplain formed. In the early Tertiary uplift again occurred, followed by a quiescent interval somewhat shorter than that of the late Cretaceous, so that only a small part of the Cretaceous peneplain was removed during the formation of the partial Tertiary peneplain. Hayes and Campbell (1894) hypothesized that by late in the Cretaceous, the Appalachian Valley southward of the New-Kanawha system constituted a single drainage system whose main trunk was a large river flowing southwestward into the Cretaceous sea, occupying approximately the present position of the Coosa River (Fig. 1). They called this hypothetical ancestor of the TR the "Appalachian River," and proposed that this river was diverted across Walden Ridge (Fig. 2) in the late Tertiary. A large westward flowing tributary of the Sequatchie in nearly the position of the present Tennessee was able to erode its headwaters eastward and divert the Appalachian River through Walden Ridge into the course of the Sequatchie River, which at that time flowed to the Gulf via a course approximately corresponding to the present-day Black Warrior River (Fig. 3). The capture was facilitated by the elevation of the Sequatchie, which was about 30 m lower than the Appalachian River, and by the more direct course of the Sequatchie to the Gulf, thus providing a steeper gradient.

Hayes and Campbell's (1894) arguments for this late Tertiary capture, however, appear dated, particularly in their reliance on peneplains. Although the concepts of erosion surfaces, together with tilting and warping of these surfaces, are sound enough, these authors were too facile in their identification of erosion surfaces and deformations of these surfaces. Some of their surfaces are now considered stripped or structural surfaces. Also, Hayes and Campbell (1894) believed that erosion surfaces essentially erase the previous landscape, allowing radical changes in drainage systems. Presently, the belief is that long-continued erosion during a tectonically quiescent interval (i.e., the latter stage of an erosion cycle), although greatly reducing relief, leaves the divides roughly in the same place (Cleaves, 1989; Costa and Cleaves, 1984; Poag and Sevon, 1989).

The evidence of Hayes and Campbell (1894) thus reduces to 1) the low, smooth divide between the TR and the Coosa River near Chattanooga (Fig. 2), with a wide valley upon the divide, suggesting to these authors that the drainages were connected in the recent past; and 2) the narrow, steep-walled nature of the Walden Ridge gorge (Fig. 2), suggesting a youthful age. Hayes and Campbell contrasted this gorge with an analogous valley, Scottsboro Valley in north Alabama, that is 10 km or more wide, suggesting to them a much greater age. Even this evidence was convincingly attacked by White (1904) and Johnson (1905b), using surprisingly modern-sounding arguments emphasizing the effect of structure on landforms. Concerning the TR-Coosa River divide, White (1904) and Johnson (1905b) pointed out that the low nature of this divide is easily explained by the bedrock, the area consisting almost entirely of dolomite and shale; the presence of a former large river is not required to explain the wide valley. Many of the northeast-southwest valleys characterizing the Valley and Ridge in this region show low, inconspicuous divides between their heads (Fig. 2), similar to that between the TR and the Coosa, even though they are occupied by insignificant streams. Concerning the Walden Ridge gorge below Chattanooga, White (1904) pointed out that the streams that run

from Walden Ridge east into the Valley and Ridge have proportionally as narrow gorges with as steep slopes as has the TR in its gorge (Fig. 2). If the Appalachian River existed and the Tennessee Valley is older than Walden Ridge gorge, these side gorges are also older, and, according to this reasoning, should be wider. Concerning the contrast between the form of Walden Ridge gorge and that of Scottsboro Valley, Johnson (1905b) pointed out that Scottsboro has a sandstone cap less than one-third the thickness of that on Walden Ridge, so that the flanks of the former valley are mainly underlain by limestone, whereas those of the latter are mainly underlain by sandstone. Differences in the forms of the valleys should be expected from lithological differences alone. He also pointed out that the cross-sectional shape of Walden gorge is very similar to that of water-gaps through hard sandstone ridges in Pennsylvania, which are attributed to structure.

Johnson (1905b) presented additional arguments opposing the capture of the TR in late Tertiary time: 1) the highly meandering nature of Walden Ridge gorge (Fig. 2) suggests a pattern inherited from a broad floodplain, rather than from a capturing stream; 2) streams along the margins of Walden Ridge and Sand Mountain have made little progress in dissection (Fig. 2), except at the gorge. If one stream was able to cut entirely through the ridge, then we should expect to find the ridge breached at other points, but it is not; 3) there is not enough elevation difference between valleys east and west of Walden Ridge to allow stream capture. In order that stream A may divert stream B from its course, A must occupy a level so much lower than B that even the uppermost headwater portions of the branch of A that effects the capture shall eventually be able to work at a lower level than that of B. That a small branch of the Sequatchie, a stream that is itself comparatively small, could work back through a high mountain barrier along a course many miles in length, and still have its headwaters low enough to capture the large Appalachian River, demands that the Sequatchie valley west of the ridge should have been much lower than the Appalachian valley to the east. There is no evidence of such a great disparity in elevations.

The proposal by Hayes and Campbell (1894) for a late Tertiary capture of the TR, diverting it into the valley of the Sequatchie anticline, thus seems unlikely. It appears that the current course of the TR through Walden Ridge was established not later than the early Cenozoic, and possibly sooner. The Appalachian River (at least the Coosa River version), if it ever existed, probably disappeared long before the late Tertiary.

Hayes and Campbell (1900) and Simpson (1900) also presented evidence from the distribution of freshwater mussels to support the stream capture hypothesis. Pleurobema, a genus of Unio, is plentiful in the TR. It is not found throughout the other portions of the Mississippi basin, but is found abundantly in the Coosa and Alabama Rivers. Simpson (1900) concluded that the upper TR at one time must have flowed southward into the Coosa-Alabama River. Johnson (1905a), however, pointed out the possibility of alternative explanations, noting, for example, that there are many reports of invertebrate organisms attaching themselves to the feet of birds, thereby possibly being spread between drainage basins. Arguments over biologic evidence of stream capture has continued to this day, in the Appalachians (Johnson, 1939, 1941, 1942; Van der Schalie, 1939; Holt, 1969) and elsewhere (e.g., Smith, 1999). Our opinion is that such evidence of former drainage connections is reasonable, but only as corroborating evidence, and is not strong enough to serve as the sole evidence of geomorphic change, unless the possibility of transfer of taxa by other than stream connections can be eliminated.

THE POSSIBLE INFLUENCE OF STRUCTURE AND LITHOLOGY ON THE COURSE OF THE TENNESSEE RIVER

Several investigators have pointed out the control of structure on the course of the TR. Hayes and Campbell (1894) speculated that after emergence of the Appalachian Mountains in the late Paleozoic, streams initially flowed to the west following the slope of the land. As the folds and faults of the Valley and Ridge became

exposed, thereby exposing beds of varying resistance to erosion, longitudinal streams south of the New-Kanawha basin, aided by the southward plunge of the fold axes, gradually developed, diverting the original streams to southern courses. During the Mesozoic, the original Appalachians eroded, until by late in the Cretaceous, nearly the whole of the Appalachian Valley southward of the New-Kanawha (Fig. 1) constituted a single drainage system whose main trunk was the Appalachian River. White (1904) suggested that all but one of the original transverse streams was captured by subsequent streams, and that exception, which still maintains its original course across Walden Ridge, is the TR.

Adams (1928) attempted to relate the history of the TR to both the Cretaceous and Paleozoic stratigraphy of the region. When Cretaceous deposition ceased, the southern Appalachian region and the adjacent Coastal Plains apparently were elevated. Following this uplift, the TR apparently attained its present position in its lower portion by extending its course over the emerging land. It made a sharp bend near Guntersville (G in Fig. 3) and developed its valley in adjustment to the underlying Paleozoic rocks as they were re-exposed. In Alabama the TR parallels the present escarpment made by the Hartselle Sandstone and follows the strike of the Paleozoics (Fig. 4). The course of the Tennessee, where it flows northward, is along the strike of the Cretaceous formations, just as the course of the Tombigbee in general is along the strike of these same rocks in eastern Mississippi and western Alabama. The TR at the northwestern corner of Alabama, where it forms the state line, flows in a relatively narrow gorge (Fig. 3) and appears to have held this position for a long time, perhaps since it established its course over the newly emerged Cretaceous deposits during the interval that preceded Tertiary deposition.

Milici (1968) noted that because many of the Paleozoic formations in central and western Tennessee presumably had been exposed by the Late Cretaceous, the physiography at that time probably was roughly equivalent to that of the present. Because the TR follows a long arcuate course around the Nashville dome (Fig. 4), Milici suggested that it may have had an initial consequent course close to its present course, probably early in the Mesozoic, and that the current course simply reflects lateral migration off the dome. He proposed that southward migration was the mechanism for entrapment of the river in Sequatchie Valley. Milici (1968) also suggested that the northward trend of the lower TR may be attributed to the following of consequent lows on the emerging Cretaceous coastal plain which reflected Late Cretaceous crustal movements or buried pre-Late Cretaceous topography. Evidence presented by Marcher and Stearns (1962, p. 1383), for example, indicates that, prior to Late Cretaceous inundation, a lowland existed between the Ozark Highlands and the Nashville dome in the approximate position of the western valley of the TR. This lowland may have been occupied by the lower reaches of the ancestral Tennessee. It was drowned during late Cretaceous flooding and is now marked by Tuscaloosa deposits. After withdrawal of Late Cretaceous and Cenozoic epicontinental seas, this lowland would have been a suitable location for establishment of the northward course of the lower TR. He attributed the present course to capture by drainage tributary to the Ohio River. Isphording (1983) pointed out that a factor favoring this northern trend might have been isostatic uplift that occurred in the continental interior and coastal plain in response to sea-level changes during the Pleistocene.

Whereas Adams (1928) speculated that a position of the TR in the headwaters of the Black Warrior River during the Cretaceous seems likely, neither Adams nor Milici (1968) saw any reason to suppose that the ancestral TR ever flowed across Mississippi. Other researchers, however, as cited below, have reported evidence of former courses of the TR in Mississippi, but during the Tertiary rather than the Cretaceous.

Although the course of the TR may be adjusted to structure and lithology, small crustal movements might also have contributed to course changes. In addition, structural control certainly does not preclude former courses much different from the present one. The best evidence of former courses would come from

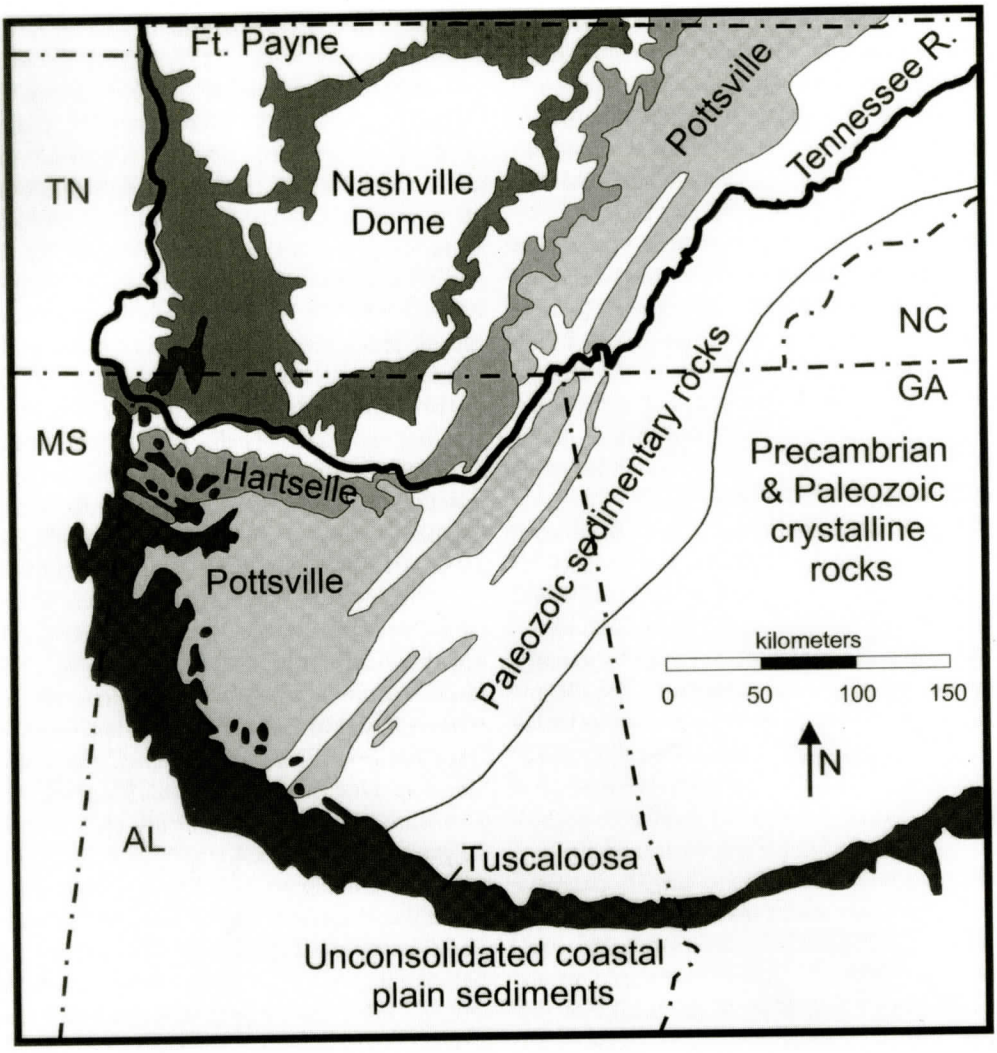

Figure 4. Important geological units and features in the vicinity of the Tennessee River.

the geographic distribution of stream deposits.

GEOLOGIC TRACERS FROM THE CRYSTALLINE APPALACHIANS

The two most easily recognized geologic tracers that indicate a crystalline Appalachian provenance are 1) distinctive suites of heavy minerals, usually rich in high-grade metamorphic minerals; and 2) metamorphic or vein quartz (subsequently these both will be referred to collectively with the abbreviation MQ, as vein quartz occurs mainly in metamorphic rocks; the fraction coming from Paleozoic sedimentary terrane is minute). The former is usually studied in the sand fraction and the latter in pebble or larger fraction.

Upstream of Chattanooga, the TR acquires abundant amounts of these metamorphic tracers from tributaries that head in the Blue Ridge province (Figs. 1 and 3). As the river crosses the sedimentary rocks downstream from Chattanooga, the metamorphics constitute distinctive markers that distinguish deposits of the Tennessee from those of streams that head in the sedimentary Appalachians. No tributaries of the TR

downstream of Walden Ridge head in metamorphic terrane. Study of geographic and temporal distribution of these tracers may help unravel the mystery of the ancestral TR. The problem of reworking of tracers will be addressed below.

Heavy Minerals

In the western part of the southern U.S., heavy-mineral suites generally can distinguish sediments from the southern Appalachians from those originating from the craton, the Rocky Mountains (via the Missouri and Mississippi rivers), and from the Canadian shield (i.e., Pleistocene glacial deposits) (Goldstein, 1942). Suites from the Appalachians are characterized by abundant high-grade metamorphic minerals, including kyanite, staurolite, and sillimanite, although the details of the suites vary. The presence of minerals from the crystalline Appalachians is to be expected, but the exact geographic and temporal distributions of these minerals may have important implications for former drainage courses.

Most heavy-mineral studies in this region have dealt with Cenozoic deposits. Grim (1936) studied heavy minerals of the Eocene sediments of Mississippi. He found that for both the Wilcox and the Claiborne, the abundant heavy minerals are uniformly present laterally across the state and vertically through the section, and therefore concluded that all the deposits of both these units had the same ultimate source area, the crystalline Appalachians.

Isphording (1983) studied the mineralogy of Miocene coastal plain sediments, and found a marked contrast in the heavy-mineral suites from Texas eastward to Florida. Texas and western Louisiana are dominated (especially in the lower Miocene) by exceptionally high zircon, apatite, pyroxene, and amphibole, derived mainly from igneous (volcanic) source areas. Upward in the section, east Texas and Louisiana Miocene sediments undergo a marked change, with a distinct decrease noted in the quantity of garnet, zircon, apatite, sphene, and especially amphibole. This reflects the decreasing effect of igneous source areas in during middle (and late) Miocene time. Instead, these sediments (as well as those from Mississippi and Alabama) became dominated by a suite typical of the crystalline Appalachians, indicating that the southern Appalachians had become the dominant source for Gulf Coast sediments.

Several studies have also been carried out on the heavy minerals of Pliocene(?) deposits. Potter (1955) studied the mineralogy of the Lafayette gravel [Pliocene(?)] deposits in the northern Mississippi embayment. He was able to divide these deposits into three geographic groups, which he considered the remnants of three coalescing alluvial fans related to the ancestral Mississippi, Cumberland-Ohio, and TR. The heavy-mineral suite for western Kentucky (i.e., the deposits associated with the ancestral TR fan) differs greatly from that of deposits from the other rivers, showing a strong metamorphic affinity suggestive of a source in the crystalline Appalachians.

Blankenship (1956) observed that the mineral suite for Pliocene(?) terrace deposits of the Mississippi in western Tennessee was markedly different from that of recent deposits of the Mississippi and also from that of older, Paleozoic formations of the continental interior. It was, instead, similar to suites from Pliocene(?) and Pleistocene deposits of the TR. He concluded that the source for the Mississippi Pliocene(?) deposits must have been sediments derived from the ancestral TR to the east. Rosen (1969) analyzed heavy minerals in the Citronelle Formation (Pliocene?) from southwestern Mississippi to the Florida panhandle. He found the heavy-mineral suite of this deposit, indicative of a metamorphic source, to be similar to that in the older underlying deposits, and therefore argued that the Citronelle represented reworked sediments from now-eroded deposits once located farther inland.

Goldstein (1942) studied the mineralogy of modern coastal sediments in the northern Gulf of Mexico. He found that the Eastern Gulf province differed greatly in its heavy mineral suite from provinces farther west, showing a relatively high content of metamorphics, reflecting a crystalline Appalachian source. Some of these minerals may be coming from the Coosa-Alabama, which has headwater branches in the

Blue Ridge and Piedmont of Georgia, but probably a substantial part are coming from the reworking of Coastal Plain sediments originally deposited there by now-defunct streams from the crystalline Appalachians.

A southern Appalachian metamorphic suite thus seems to be ubiquitous in the eastern Gulf Coastal Plain, in Eocene and younger deposits, indicating an ultimate source in the crystalline Appalachians, but leaving unanswered the question of the immediate source of the minerals.

Metamorphic Quartz Gravels

The great bulk of gravels on the Gulf Coastal Plain consist of chert, with MQ making up only a tiny fraction of the deposits. Clasts of MQ have been the most widely cited evidence for former courses of the TR. MQ gravels have the advantage over heavy mineral suites that they can be seen readily in the field. Also, the MQ gravels resist weathering more than all but the most-resistant heavy minerals.

MQ gravels have the same problem as heavy minerals, in that the ultimate source is readily identified but the immediate source is more difficult to ascertain due to reworking by streams. It may be possible to alleviate this problem for certain time ranges or geographic locales. The first restriction is that MQ associated with the Coosa-Alabama river system cannot be assumed to be derived from a former course of the TR, as the Coosa has headwaters in the Piedmont and Blue Ridge of Georgia (Figs. 1 and 4). Gulf rivers west of the Coosa-Alabama (Fig. 3), however, must ultimately have derived their MQ from the headwaters of the TR or its ancestor.

The two most likely immediate sources of MQ other than the TR are the Pennsylvanian Pottsville-equivalent conglomerates and the Late Cretaceous Tuscaloosa Formation (Fig. 4). Pottsville outcrops in the Appalachians display MQ clasts no larger than a large marble. It is possible that up-section beds long since removed had larger clasts, but it seems unlikely that Cenozoic deposits with MQ clasts larger than this were obtained from the Pennsylvanian formations.

Large MQ clasts do occur in Tuscarora deposits. However, the Tuscaloosa varies greatly from place to place in the amount of MQ clasts present. Marcher and Stearns (1962), for example, found that in Tennessee, lithologies of the Tuscaloosa differ from west to east. The western facies, typical of the formation in most parts of the western Highland Rim, consists of poorly sorted chert gravel with a small percentage of sandstone pebbles. Most of the gravel is Devonian chert, apparently from the Pascal arch, an eastward-sloping extension of the Ozark dome that extended into west Tennessee as far as the Tuscaloosa sea. A smaller amount of gravel is Mississippian chert of local origin.

The eastern facies, located near the eastern erosional limit of the Tuscaloosa in Tennessee, is similar to the western but with the fines winnowed out and MQ sand and pebbles added. Locally there are beds and pods of well sorted, heavy-mineral-bearing MQ sands. Beds of well sorted gravel that locally contain an abundance of MQ pebbles also occur. Marcher and Stearns (1962) suggest that the MQ and Blue Ridge-type heavy minerals in this eastern facies may have been derived from Pennsylvanian sandstone and conglomerates that cropped out north and south of the Pascola arch. The abrupt appearance of abundant MQ pebbles in the eastern facies of the Tuscaloosa coincides with an equally abrupt improvement in sorting. This coincidence suggests that MQ pebbles were brought into the Highland Rim area from the north or south and that they were deposited under different conditions than was the main mass of chert gravel.

Most of the Tuscaloosa Formation is believed to be of nonmarine origin, although the eastern facies is believed to be partly marine in origin, the exotic components having been swept in by longshore currents. The Tuscaloosa generally is thin, preserved only as caps on outlying hills. In Wayne and Hardin counties, in the southwestern part of the Highland Rim (WA and HN in Fig. 3), the gravel is intermixed with residuum derived from the underlying rocks. The greatest thickness is in parts of southwestern Wayne County, where 150 ft of gravel has

been reported (Marcher and Stearns, 1962).

In western Alabama, Monroe and others (1946) reported that only the Gordo Formation of the Tuscaloosa Group contained substantial gravel. The gravel consists predominantly of chert pebbles derived in large part from the Mississippian limestones of northern Alabama and southern Tennessee. Locally, however, as many as one fourth of the pebbles are MQ, perhaps derived from the much nearer conglomerates of the Pottsville formation. The Gordo Formation has been recognized as far east as Chilton County and as far north as the TR, where it includes the large gravel deposits mined extensively in the Iuka area in Tishomingo County, Mississippi (IU in Fig. 3).

These studies show that locally the Tuscaloosa may have a relatively high (i.e., up to 25%) content of MQ clasts, but that, overall, the Tuscaloosa is composed primarily of chert, so that reworking of this unit by streams is unlikely to produce a deposit high in MQ clast content. Marcher and Stearns (1962) suggest that, at least within the area covered by the western facies of the Tuscaloosa, there should be little difficulty in differentiating remnant Tuscaloosa deposits from TR deposits on the basis of MQ clast content. In fact, the authors point out that in this area, the general absence of MQ pebbles, together with the absence of waterworn ironstone pebbles, are the main criteria for distinguishing the Tuscaloosa from much younger terrace gravel. In the eastern facies, however, differentiation of the two types of deposits may be difficult.

Since the percentage of MQ clasts deposited at a given time is a function of the rate of chert deposition as well as that of MQ deposition, it is of interest to ascertain the factors affecting the former. The rise of the Pascola arch in the Late Cretaceous produced a flood of Devonian chert (Marcher and Stearns, 1962). A much later event which may have been important is the breaching of the Nashville dome (Fig. 4). Reesman and Godfrey (1981), based upon erosion rates calculated from dissolved loads measured in streams in the Central Basin and the Highland Rim of Tennessee, estimated that the Fort Payne Chert would have begun to be breached about 5-6 Ma ago, leading to a flood of cherty gravels in the Gulf Coastal Plain and Mississippi embayment (Self, 1993). Although this inference from erosion rates provides a hypothesis that can be tested by future dating, it is rather speculative and can be used to establish only a very approximate age. The problem is, that although the reconstruction of the Nashville dome based on denudation rates may be reasonable with respect to the rock volume, it minimizes the probable irregularity of fluvial erosion. Deep fluvial erosion may have cut down into the Fort Payne, supplying much debris from this formation while most of the formation was still covered by younger rocks. Thus, debris from the Fort Payne might have been voluminous much earlier than 5-6 my ago. Russell and Parks (1975) attributed the sudden influx of chert to a completely different cause. According to them, in southern Tennessee when the river was flowing at what are now elevations of 700 ft and above, the TR was carrying mainly MQ and sandstone gravels. As it migrated westward and cut below what is now the 700-ft level, the Tuscaloosa Formation was unroofed, releasing a flood of chert to the TR.

MQ clasts may thus serve as indicators of immediate, as well as ultimate, sources of late Cenozoic fluvial deposits if attention is confined to larger clasts and/or to areas where MQ-rich Tuscaloosa deposits do not crop out up gradient.

HIGH TERRACES AND THE STUDY OF TENNESSEE RIVER DEPOSITS

One problem in studying abandoned river courses of the TR, or even in evaluating possible scenarios of drainage evolution, is our lack of knowledge concerning how TR terraces change with age and how long they survive. One way to improve this knowledge base is to study high-level terraces near present-day river courses, where the terraces can be associated with specific streams with near certainty, and their ages roughly estimated from their elevation above modern river level (AMRL), assuming incision rates are approximately known. Understanding of these terraces can then help to

evaluate hypothesized stream deposits or courses far from modern streams.

Such a study has been carried out along a reach of the New River in southwestern Virginia (Mills and Wagner, 1985; Bartholomew and Mills, 1991) (Fig. 1). The New River heads in the Blue Ridge province of North Carolina, and carries a large load of MQ gravel. In the study area in southwestern Virginia, tributaries head only in the Valley and Ridge, and thus carry no MQ gravel. Therefore, high terraces of the New River can readily be identified and distinguished from those of other streams by the presence of rounded MQ gravels. At the time this study was done, incision rates of the New River could only be estimated. However, a rate of 27.3 m/m.y. has since been measured by Granger and others (1997) by means of cosmogenic isotope dating of MQ clasts in riverside caves. Ages of terraces can now be approximately dated in millions of years by dividing terrace elevations AMRL by 27.3.

Mills and Wagner (1985) studied erosional and weathering characteristics of the New River terraces. They found that the original surface of terraces less than 25 m AMRL is largely intact. The surface of terraces 25-50 m is highly dissected, but remnants of the original surface remain. Above 50 m, little or no original surface remains; terraces have been transformed into rolling, irregular topography, with numerous prominent sinkholes in areas underlain by carbonates. Deposits survive much longer over a limestone substrate than over a shale one (Houser, 1981). Reworking of alluvium by hillslope processes is very common on high "terraces", and on terrace remnants higher than 100 m AMRL, there appears to be little in situ alluvium remaining. In deposits above 50 m AMRL, most crystalline rock clasts are decomposed, and in the sand fraction, most feldspar grains are gone, as are the less-resistant heavy minerals.

Despite the intense weathering and erosion of the high-level deposits, colluviated remnants of terrace deposits, recognizable by their MQ clasts, occur as high as 300 m AMRL, translating roughly into an age of 11 Ma. Deposits 185-275 m AMRL (6.8-10.1 Ma) are widespread, although they occur in small isolated patches (Bartholomew and Mills, 1991). Some of these deposits are very near the modern river, whereas others are located as far away as 11 km and describe, in some cases, river courses much different than the modern one.

The work on the New River in southwestern Virginia shows that fluvially deposited MQ gravels remain as testimony to the former presence of a stream for long time periods, even though original terrace surfaces are gone by 2 my and original deposits are almost completely reworked by slope processes by 4 my. This work points out the importance of determining both the locations and elevations AMRL of the deposits, and provides a model that could be used for study of the TR, although incision rates and survival times of terrace surfaces and original deposits may differ. There are certainly more difficulties involved with study of TR than New River terraces. On high terraces of the New River, most of the deposits are reworked, but generally they can be readily identified as coming from the New River by their MQ content. This is not the case for gravels of the TR, where MQ may come from Paleozoic or Mesozoic sources, as well as from late Cenozoic ones. This problem is more manageable for terraces near the modern Tennessee, but becomes very serious for deposits of possible former courses of the Tennessee tens or even hundreds of kilometers from the present river, especially where they overlie Gulf Coastal Plain deposits.

Few detailed terrace studies have been carried out along the TR. Delcourt (1980), provided a basic study of the Little Tennessee River, a tributary of the TR that heads in the Blue Ridge province (Fig. 3). He included high terraces, but did not include information on still higher deposits above the highest preserved terrace surfaces. Archaeological studies have dealt only with low TR terraces. In the upper reaches of the TR, there have been some observations on high-level gravels. Swingle (1959), for example, described high-level gravels east of the river near Cleveland, Tennessee (CL in Fig. 3), ranging up to 700 ft (213 m) AMRL, although the highest ones consist mainly of quartzite rather than MQ.

One test of the Walden Ridge late-Tertiary capture hypothesis might be to compare heights of gravels upstream and downstream of the ridge. If the upstream reach is substantially older than the downstream, gravels might be expected to occur at higher elevations AMRL upstream than downstream. Hayes and Campbell (1894), for example, claimed that they found MQ gravels upstream of the ridge up to 250 ft (76 m) AMRL, but downstream only up to 150 ft (46 m). However, Johnson (1905b) claimed to have found MQ gravels up to 265 ft (81 m) AMRL downstream of the ridge, and we have found them at similar heights. Thus, the data do not support the capture hypothesis.

For the lower reaches of the TR, aside from the deposits near the junction with the Ohio studied by Potter (1955), the most extensive terrace deposits occur just north of the bend at Pickwick Dam (PD in Fig. 3), where the TR assumes its northward course across Tennessee, with few terraces occurring north of this area. These deposits have been mapped by Russell (1964, 1967, 1968), Russell and Wilson, 1970; Russell and others (1972), Wilson and others (1971, 1982), and have been discussed by Russell and Parks (1975). Self (2000) divided these deposits into five levels, with heights AMRL of the treads ranging from 80-140 ft (24-43 m) for the lowest to greater than 360 ft (110 m) for the highest. (Self reported tread heights in altitudes, but we have transformed these into heights AMRL by subtracting the approximate pre-dam level of the TR from the altitudes). Thickness of the deposits ranges from 40 to 60 ft (12-18 m). Only the lower two terraces, those with treads with heights AMRL of 160 ft (49 m) or lower, show well preserved surfaces.

Dating these deposits from their heights AMRL is impeded by the lack of a measured incision rate. The rate is probably lower than the previously cited 27 m/my of the New River in southwestern Virginia, owing to lower relief along the lower TR. However, a compilation of incision rates in the eastern United States (Mills, 2000) suggests only a weak correlation between relief and incision rate, so that the TR rate is probably not greatly less than that of the New River rate. If we assume that the TR incision rate is, say, between one half and one times that of the New River rate (range of 13.5 - 27 m/my), the age of the highest terrace would be 4.1 to 8.2 Ma. Self (2000), by means of lithology counts, found that the clasts on the highest terraces are largely MQ, whereas those on lower terraces are largely chert. He attributed this change from predominantly MQ to predominantly chert to the breaching of the Nashville dome, estimated by Reesman and Godfrey (1981) to have begun about 5-6 Ma. The incision rates estimated here support this timing.

Self (2000) also found that both terraces and the underlying Claiborne Formation (middle Eocene) in the Hatchie River west of his study area showed a high content of MQ clasts. He therefore suggested that the ancestral TR flowed through the Hatchie River valley prior to the breaching of the Ft. Payne Formation, and that it may have followed that course at least since the Eocene. He noted that additional support is provided for this concept by the width of the upper Hatchie valley, which suggests that the present Hatchie River is underfit, and by the alignment of the Hatchie valley with the axis of the TR before it turns to the north.

Self's (2000) hypothesis could be tested by correlating the terraces in the two valleys. No absolute dates are available for terraces, nor is the stream incision rate for either stream. However, correlation can be attempted by comparing the elevation AMRL and degree of preservation of terraces. Parks (1968, 1992) studied terraces along the Hatchie River valley near Hebron, Hardeman County, Tennessee (H in Fig. 3). He found that four levels of terraces occur on the southwestern valley wall, their treads ranging in elevation AMRL from 14 to 60 m. Only the lowest terrace, at 14 m AMRL, is well preserved. This contrasts with the TR, where, according to Self (2000), the highest terrace is 110 m AMRL, and the highest well preserved terrace is at 49 m AMRL. These discrepancies, particularly the 14 m vs. 49 m elevation of the highest preserved terrace, suggests that the terrace sequence between the two valleys are not correlative, unless the incision rates differ by more than a factor of 3, which seems unlikely

A further attempt at chronology can be made

by using a correlation suggested by Saucier (1987). He named and described four lower-level terraces along five West Tennessee streams that drain to the Mississippi. From highest to lowest these are the Henderson terrace, the Humbolt, the Hatchie, and the Finley. The lower three were recognized along the Hatchie river, but only the Hatchie terrace was recognized on the Hatchie River as far upstream as the vicinity of Hebron. This corresponds to the low terrace of Parks (1992). Saucier (1987) postulated that the Hatchie is the stratigraphic equivalent of the Prairie Terrace, which is the Prairie Complex of Autin and others (1991).The latter authors report that age estimates for the Prairie Complex range from Sangamon to late Wisconsin. (These estimates are compatible with Parks' [1992] suggestion that all the Hebron terraces are Pleistocene.) If we assume a Sangamon age (on the order of 120 ka), then if the Hatchie terrace does indeed correlate with the 49-m-high TR terrace near Pickwick, the implication would be that the highest TR terrace, at 110 m AMRL, is probably less than 1 Ma. As an age this young seems unlikely, the inference must be, once again, that the highest preserved terraces at the two locales probably are not correlative. Hence, terrace correlation provides no support for Self's (2000) interpretation of the Hatchie as the former course of the TR, as the Hatchie River deposits appear to be substantially younger than the TR deposits near Pickwick. However, the evidence has too many uncertainties to dismiss his hypothesis. Much more study of TR terraces, particularly dating, is required before significant progress will be made on the drainage history of the TR.

THE UPLAND DEPOSITS

Sands and gravels that unconformably overlie Paleozoic to Miocene formations and underlie Quaternary formations are widespread in the Gulf Coastal Plain and the upper Mississippi embayment. These coarse deposits contrast sharply with the fine sands, silts, and clays that characterize the Quaternary and much of the Tertiary in the Gulf Coastal Plain (Self, 1993). Names for these deposits include the Citronelle of the Gulf Coast, the Lafayette gravels of the Upper Mississippi embayment (Autin and others, 1991, have pointed out that the name Lafayette is now obsolete, but it will be used herein), and the Upland Complex in the Lower Mississippi Valley (Autin and others, 1991). The latter name, modified to Upland deposits, will be used here for all these deposits. The gravel fractions are dominated by rounded, iron-stained, honey-colored chert, whereas the sand fractions are largely MQ. Potter (1955) proposed that the gravel comes mainly from the sedimentary Appalachians and the sand mainly from the crystalline Appalachians. The deposits are generally assumed to be Pliocene or early Pleistocene in age, although age control is poor. Because these deposits commonly contain crystalline Appalachian suites of heavy minerals, and often some MQ gravel as well, they may be related to the former drainage of the TR.

Although some Upland deposits include high-level terrace deposits associated with specific streams, generally these deposits are located away from large modern streams so that their origin is more uncertain. Shaw (1918) suggested four possible origins of the Upland deposits:

1) they were deposited by floodwaters of Pleistocene glaciers;

2) they were laid down during a marine submergence of the Coastal Plain;

3) they were produced by stream deposition induced by broad uplifts of the Appalachians;

4) they are for the most part simply weathered portions of older, underlying formations of the region, with a small part being made up of material of other kinds.

The first two can now be largely eliminated as possibilities, although ice-rafting of rock debris by streams during glacial climates may have been significant. The second two, however, remain viable. Origin 3) seems to be the most broadly accepted today. The idea is that increased uplift of the Appalachians during the Pliocene increased the load of the streams, and that this load was too great to be carried across the coastal plain to the sea. As the uplift progressed, the deposit was spread farther and farther toward the sea. The basic evidence for this interpretation is that the deposits roughly form

an arc around the Appalachians. Self (1993) thought this uplift to be concentrated in the vicinity of the Nashville Dome, which Reesman and Stearns (1989) hypothesized to have undergone extensive isostatic uplift during late Tertiary and Quaternary time owing to rapid erosion following the breaching of the capping Ft. Payne Formation. The coarse texture of the deposits, in contrast to the fine texture of most of the Tertiary coastal plain deposits, suggests a unique depositional regime (Self, 1993). Autin (1991) suggested that the deposits represent erosional remnants of a once regionally extensive blanket, commonly capping hilltops or on well-dissected interfluve ridge crests. Potter (1955) and Self (1984, 1993) suggested that the Upland deposits were laid by braided streams.

Shaw (1918) has given the most complete discussion of origin 4). Working in northern and central Mississippi, he opined that 75% to 90% of the supposed "Lafayette" deposits appear to be material from the underlying formation subjected to weathering and reworked by colluviation. As evidence, he noted that the supposed formation shows little detailed or consistent relationship to altitude, topography, or geologic features. Further, the surficial material often differs from the intact pre-Pliocene formation beneath only in being more reddened and otherwise weathered. In addition, the unconformity that should occur beneath the deposit either cannot be found or cannot be agreed upon.

On the other hand, Shaw (1918) found that clasts on or within a few feet of the surface of the Upland deposits in many cases are much larger than any found in the underlying formations, or, in fact, in any other formation in the region. Clasts may rarely be as large as boulder size; Mellen (1939) reports a boulder weighing 905 lbs. These large clasts evidently are remnants of strata now worn away, and the question arises of whether these strata belonged to a single formation or several formations, some or all of which are now represented by the Upland deposits. Shaw argued that if the clasts had been let down from a single formation they should show some relationship to altitude and surface features, but they do not, and so derivation from multiple formations seems more likely. As a source of the large clasts, Shaw (1918) pointed out that the landward part of many coastal-plain deposits is likely to be coarser than the seaward part, and it is thus probable that the now-eroded landward parts of the formations in the study area once contained clasts coarser than those found in the present-day outcrops.

Shaw (1918) also addressed the question of when the clasts moved from their positions in the last formation they were part of to their present position. He argued for a Pliocene age as follows. The present rate of erosion is about 100 ft/my (an estimate that remains today the right order of magnitude). The clasts are likely to have been let down many feet, for they are found only in colluvium. If they had been lowered much less than 100 ft, one might expect to find remnants of the beds from which they were derived, but no such remnants have been found. The clasts have not been let down many hundreds of feet, however, for they are fairly evenly distributed and are almost as common on divides as elsewhere.

The other 10% to 25% of "Lafayette" deposits, which are clearly different from the underlying formations and separated from them by an unconformity, Shaw (1918) attributed to terrace deposits, and noted that most of them are in the vicinity of large streams. These deposits have been modified in a way similar to the other Upland deposits. As an example, he pointed out gravel deposits near the junction of the Tombigbee and Black Warrior rivers, which he attributed to remnants of several different terraces that stood at different heights, and hypothesized that the large bodies of gravel have been let down, with lateral shifting, as the underlying materials eroded.

Although the consensus today favors origin 3) for the Upland deposits, certainly Shaw's (1918) account seems reasonable, and should be kept in mind as a possible explanation of at least some of the Upland deposits.

Concerning courses of the TR, although origin 3) considers many coarse gravels to be part of an eroded sediment blanket surrounding the Appalachians, it does not eliminate the use of gravel deposits as evidence for former stream courses. It does imply, however, that the simple

presence of such deposits does not, by itself, demonstrate the existence of such courses.

INFERRED FORMER COURSES OF THE TENNESSEE RIVER

Previous Work

Most studies of Upland deposits were not concerned with specific relict stream courses. An exception is that of Potter (1955), discussed earlier, who associated part of the Upland deposits of western Kentucky with the ancestral Tennessee. His evidence seems to imply that the lowermost reach of the TR was established by Pliocene, which implies that upstream reaches of the TR most likely had been established by that time as well. Of terrace studies, only that by Self (2000) bears on previous TR courses. Based on the presence of abundant MQ on terraces and in the Eocene cropping out in the Hatchie Valley, as well as other evidence discussed in the previous section, Self suggested that this valley had been the course of the TR as early as Eocene, and that this course had been abandoned by the time the TR began carrying abundant chert gravel at about 5-6 Ma.

Most reconstructions of former courses of the TR have been based on gravel deposits, particularly those rich in MQ. The only such studies to deal with pre-Tertiary courses are those by Monroe and others (1946) and by Conant (1964). Monroe and others (1946), during mapping of Cretaceous units in Alabama, found that the sediments in nearly every formation were somewhat coarser near the present Black Warrior River, suggesting to them that the Black Warrior may follow the approximate course of a Cretaceous stream, possibly the ancestral TR. Conant (1964) discussed this subject in more detail. He noted that a water well at Brent, Alabama (B in Fig. 3), penetrated Cretaceous gravel and other sediments in a paleochannel in Paleozoic bedrock that was about 40 feet deep. Thus, during Cretaceous time a major stream may have had a course similar to that of the present Cahaba River (Fig. 3) in the Brent area, and was entrenched at least 40 feet in the Paleozoic rocks. About 12 miles downstream from this well, an exploratory well penetrated 30 feet of gravel at the base of the assumed Vick Formation (?). The Cahaba River is directly in line, geographically and structurally, with the TR above Chattanooga. The Black Warrior River is directly in line with the TR in its southwest course along the breached Sequatchie anticline. Conant (1964) suggests that in Cretaceous times the TR continued southwest from Chattanooga to the course of the present Cahaba River, and that the Sequatchie River flowed into the course of the present Black Warrior River. Such courses would explain the greater abundance of gravel in the Cretaceous sediments near the present rivers. Neither Monroe and others (1946) or Conant (1964) discussed gravel lithologies.

Concerning Tertiary courses, Grim (1936), based on heavy-mineral data, suggested that sediments from the crystalline Appalachians accumulated in the northeast-central part of Mississippi, probably in the form of a huge delta into the embayment fed by a large river, perhaps the Appalachian River of Hayes and Campbell (1894) during Midway and Wilcox time in the Eocene. On the other hand, during the younger Claiborne time, there is no evidence of such a river or delta in this location. As an explanation, Grim favored the idea that capture of the Appalachian River took place approximately at the beginning of Claiborne time. He offered no evidence of specific courses, however.

Most speculations concerning former courses of the TR have dealt with the late Tertiary.

Hayes and Campbell (1894) hypothesized that at the close of the Cretaceous a small river flowed westward across northern Alabama and emptied into the sea in the northeast corner of Mississippi. During the late Tertiary one or more of the head branches of this stream then captured some eastern TR course that had been going to Mobile Bay, and upon the withdrawal of the sea this stream then followed the course of the Big Black (Fig. 3) to the Mississippi. Shaw (1918) agreed with the concept of a former course in Mississippi, but thought a course down the Big Black in Pliocene time unlikely, for no high terraces occur along this river

and there is no abandoned valley between the two streams. He thought a more likely route to be down the Tombigbee, for along this river there are somewhat extensive high terraces, particularly in the vicinity of West Point, Mississippi (WP in Fig. 3), and to the southeast in Alabama. If this course existed, some parts of it have not been found.

Shaw (1918) also considered the Pearl River in Mississippi (Fig. 3) to be underfit to its valley; as an explanation he suggested that the western headwaters of the Pearl were captured by the Big Black River, perhaps in late Pliocene time. Perhaps another possibility might be, however, that the Pearl was once occupied by the lower end of the ancestral TR or other stream from the Appalachians.

Brown (1967) hypothesized that a large river flowed southwesterly through southern Mississippi in Pliocene times, based on the discontinuous gravel-defended ridges extending along this trend, particularly in Scott, Lincoln, and Franklin counties (SC, LI, and FR in Fig. 3). The thick superficial deposits that underlie the ridges, which contain very coarse gravel, have been mapped as Citronelle. He suggested that many deposits considered to be Eocene or even Cretaceous might in fact prove to be part of this hypothesized river. The alignment of the gravel-defended ridges in southern Mississippi suggests a flow direction compatible with the undated and somewhat anomalous gravel deposits of Tishomingo, Choctaw, and Winston counties (TI, CT, and WI in Fig. 3). Brown did not discuss the lithologies of the gravels.

Isphording (1983) reported a local anomaly in the heavy mineral suite at a site south of the town of New Augusta in southeast Mississippi (NA in Fig. 3) which led him to hypothesize a former river course. Samples from numerous drill holes from this location showed that significant amounts of hornblende, epidote, and garnet were present in the Miocene Hattiesburg clay. Samples from the same formation to the east and west of this location, however, did not contain these minerals. Similarly, these minerals are completely absent in contemporaneous sediments eastward in Alabama. As this suite is contained within sediments having an obvious fluvial origin, he suggested that they may have been transported into the area by some ancestral river system originating in a metamorphic rock terrane. Isphording suggested that an ancestral TR may have flowed southwesterly across the state of Alabama into eastern Mississippi, and discharged into the Gulf of Mexico in eastern Louisiana.

The above course reconstructions are not very satisfying. In the first place, many of them do not address adequately the content of tracers from the crystalline Appalachians. Secondly, they deal mostly with deposits in a relatively restricted area, avoiding the question of how the stream got to that location and where it went afterwards. Related to this problem are the missing parts of hypothesized courses, referred to by Shaw (1918). Although part of a former river course may have been down the valley of a present-day stream, there are bound to be divergences of old and new courses where the old stream joined the present one. For abandoned courses of Pleistocene or Pliocene age, some remnants should remain.

New Data Bearing on Former Courses of the TR

Figure 5 shows locations of surficial gravels that contain high fractions of MQ clasts. We think that many of these are remnants, commonly colluviated, of high-level fluvial deposits of Plio-Pleistocene age that potentially represent the TR. These deposits lie on the surface, and in many cases overlie deposits composed mainly of chert. Although we cannot prove beyond a doubt that these gravels are not remnants of pre-Pliocene formations, they constitute data that should be considered in any reconstruction of former river courses. The altitude ranges of the deposits (Fig. 5) may bear on the relative ages of possible previous courses of the TR, although the resistance of the underlying formations plays an important role in determining the rate of gravel lowering. For example, the gravels on the Hartselle Sandstone have probably been let down much less than those overlying the chalk belt.

The cluster of sites near A in Figure 5 include

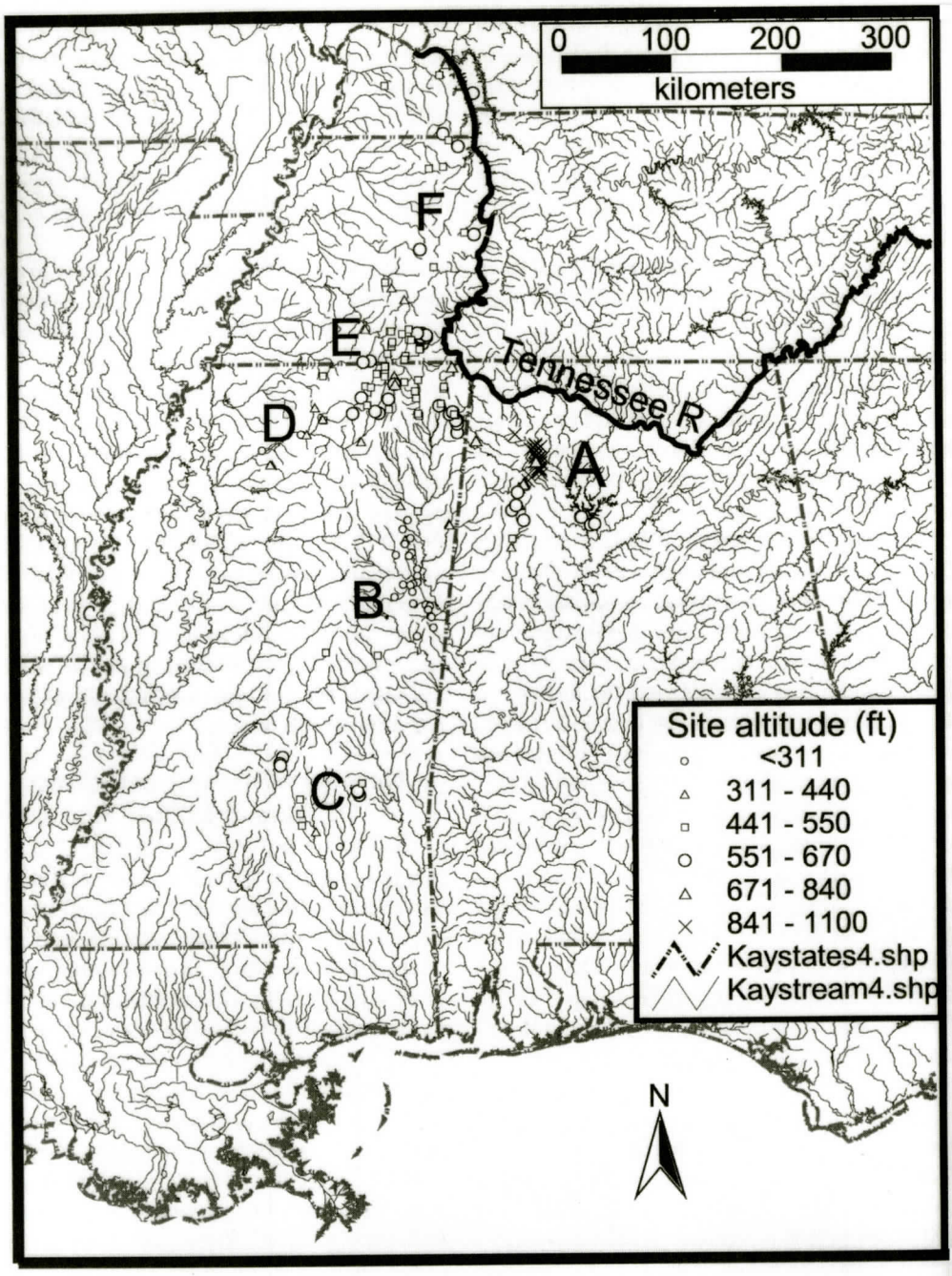

Figure 5. Map showing locations of sites with high MQ gravel contents. Six altitude intervals are shown.

those at the highest observed altitudes, which overlie Mississippian Hartselle Sandstone. These deposits presumably are the oldest, and might even be as old as Miocene. Because the downdip migration of the TR here is south, these gravels weren't simply left behind as the river migrated. Rather, they suggest an older course of the TR to the south, although its route to the Gulf is unknown. In the southern part of this group, along the Sipsey River valley (Fig. 3), are voluminous deposits with some of the largest MQ clasts west of the Blue Ridge. Because these sites are much lower and therefore probably younger than those on the Hartselle, they may reflect a later course of the TR. These deposits contrast with those of the Luxapalila River Valley (Fig. 3), a short distance west of the Sipsey. This valley is dominated by chert gravel, thereby being typical of coastal plain streams. The easternmost sites in group A suggest a third alternative, a course to the southeast into the Black Warrior River basin. (Although the Black Warrior River is in line with the Sequatchie Valley anticline, a former route from the Sequatchie to the Black Warrior seems unlikely, as no MQ gravel has been found in the headwaters of the latter.)

The cluster at B, mainly on the west side of the Tombigbee River, has the lowest altitude of any of the clusters. Even though deposits with lower altitudes are generally younger than those with higher altitudes, these deposits are not younger than deposits along the present northern course of the TR. They may be former TR deposits that have been reworked by the Tombigbee, or TR deposits that have been let down by the rapid solution of the underlying chalk. The cluster at C might be a continuation of either cluster at B or C. It is close to, but still north of, the anomaly near New Augusta, Mississippi, reported by Isphording (1983) and attributed to a former course of the TR. We have not examined gravel lithologies in southwestern Mississippi, and therefore cannot evaluate Brown's (1967) hypothesis of a former TR courses in this area.

Cluster D suggests a former westerly course of the TR to the Mississippi River. Cluster E is associated with a fan-like deposit near the Mississippi-Tennessee border, discussed below, and also with the headwaters of the Hatchie River. If the Hatchie River does indeed represent a former course of the TR, as suggested by Self (2000), then it seems reasonable to infer that this course came after the course associated with cluster D. Cluster F shows MQ clasts west of the present Northern Valley of the TR, suggesting that the TR at some time occupied positions to the west of its present lower reach.

DISCUSSION AND CONCLUSIONS

Isphording (1983) suggested the possibility of northward diversion of drainage owing to crustal tilting caused by isostatic adjustment due to sea-level change during the Pleistocene. Crustal loading by the ice sheet covering the Midwestern U.S. is another possible cause. Presumably the first large glaciation would have been the one responsible for the diversion, probably the same glaciation that produced proglacial lakes extending far from the glacial margin, obliterating the Teays River, and creating the modern Ohio River. Granger and Smith (1998) have dated the proglacial lake sediment, by means of burial dating with cosmogenic isotopes, at 1.13 Ma. Although the most dramatic effects took place in the immediate vicinity of the ice sheet, farther south the crustal subsidence would have tilted the surface northward, thereby increasing the energy of streams flowing north and promoting capture of streams with different flow orientations. This effect might have diverted the lower reach of the TR to its junction with the Ohio.

Tilting to the north should have reduced the energy of south-flowing streams, potentially producing fan-like deposition where the slope became sufficiently reduced. Such a fan, now dissected, appears to be present in northeastern Mississippi and adjacent Tennessee. For example, in southernmost Alcorn County, Mississippi (AL in Fig. 3), extending over a wide area in the Rienz community, is a deposit of MQ pebbles in silty clay. Evidence of stream reversal in this area is also suggested by drainage maps of north-flowing streams in this area. A number of these streams have headwater tributaries that

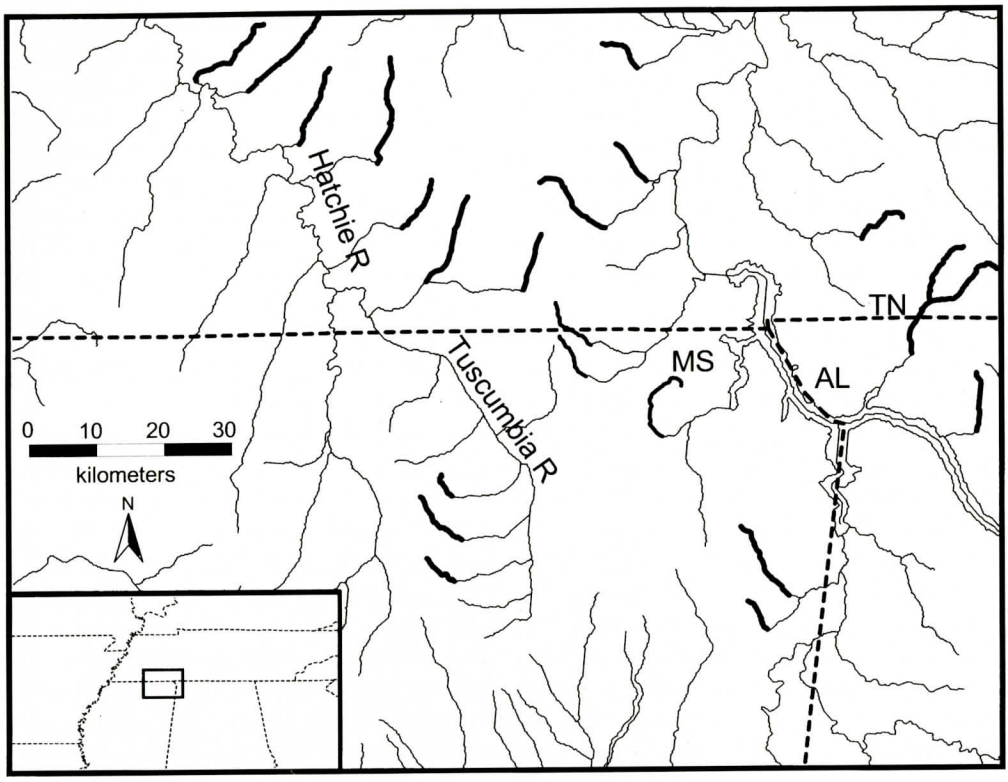

Figure 6. Map showing southward-flowing headwater tributaries, suggestive of stream capture, in northeastern Mississippi and adjacent Tennessee. Southern-flowing tributaries are shown by heavy solid lines.

flow south (Fig. 6), indicating a former drainage system flowing to the south.

Two objections that might be made to a stream capture resulting from isostatic adjustment of the ice sheet are that 1) the ice sheet has come and gone repeatedly during the Quaternary. Therefore, would not the stream course be diverted back to a southward course during each postglacial isostatic rebound? and 2) an age of 1.13 Ma seems too young, for if capture had taken place that recently, one might expect more evidence of the former stream course to remain. The first objection can be answered fairly readily - once a capture has taken place, hydrologic changes occur which provide the stream ability to persevere in its course despite later unfavorable tilting of the land surface. (The increased discharge resulting from the capture allows the stream to have more energy at a lower slope.) Concerning the second, the major question is just how well valley topography persists over time. The Teays Valley in West Virginia (Fig. 1), for example, is easily recognized despite a probable age of 1.13 Ma (Granger and Smith, 1998). Yet, there appear to be no comparable abandoned valleys in western Tennessee. The western Hatchie River Valley, discussed above, might be a candidate, but what is missing is a valley connecting the present course of the TR to the Hatchie - the divide between them seems more prominent than what should be expected in such a relatively short time interval. Self's (2000) suggestion of a minimum capture age of 5-6 Ma thus appears more compatible with the present topography. Such an age would eliminate ice sheets as the cause of diversion.

Future work should include mapping and attempted dating of terrace and other high-level deposits, more study of deposit lithologies, and

attempts to trace possible former river courses over longer distances. Terrace mapping is particularly important in order to determine the heights of river deposits above the level of the modern river. Such work would be especially useful if accompanied by dates that would allow estimation of incision rates. At present the most feasible method of dating deposits of Plio-Pleistocene age in this environment appears to be the cosmogenic isotope burial dating. One approach might be to use MQ clasts in riverside caves along the TR in reaches where it flows on Paleozoic bedrock, as done by Granger and others (1997) on the New River in Virginia. Another might be to use MQ clasts buried beneath gravel deposits.

More lithology counts of gravel deposits, such as those by Kaye (1974) and Self (2000) should be carried out. An understanding of local variation and regional trends in MQ percentages, for example, would be useful for narrowing the possibilities of former river courses. Another valuable contribution would be quantitative comparisons between MQ clast size and abundance in pre-Pliocene deposits and those in Plio-Pleistocene deposits. Such comparisons would aid in determining the immediate source of high-level Plio-Pleistocene deposits. A desirable characteristic of future studies would be the consideration of longer reaches than have most previous studies. An anomaly at one location suggesting a former course of the TR, for example, has much more significance if it can be linked with a possible upstream or downstream continuation of the course. In addition, studies comparing terraces of upstream and downstream reaches of the TR would be very useful.

We think that one possible source of confusion in the study of MQ deposits is that some late-Cenozoic MQ-rich sites may have been misidentified as exposures of early to middle Tertiary or Cretaceous formations. A re-examination of MQ-rich exposures in these older formations may show that some of them are actually late Cenozoic.

In conclusion, a review of previous work on the geological history of the Tennessee River shows many suggestive findings, but few that are substantial. The problem of deciphering the complex and confusing evidence is inherently difficult. However, it is also true that intensive, detailed studies of this topic are still few in number.

REFERENCES CITED

Adams, G. I., 1928, The course of the Tennessee River and the physiography of the southern Appalachian region: Journal of Geology, v. 36, p. 481-493.

Autin, W. J., Burns, S. F., Miller, B. J., Saucier, R. T., and Snead, J. I., 1991, Quaternary geology of the Lower Mississippi Valley, in Morrison, R. B., ed., Quaternary nonglacial geology:Conterminous U. S.: Boulder, Colorado, Geological Society of America, The Geology of North America, v. K-2, p. 547-582.

Bartholomew, M. J., and Mills, H. H., 1991, The course of the New River: its late Cenozoic migration and bedrock control inferred from high-level stream gravels in southwestern Virginia: Geological Society of America Bulletin, v. 103, p. 73-81.

Blankenship, R. R., 1956, Heavy-mineral suites in unconsolidated Paleocene and younger sands, western Tennessee: Journal of Sedimentary Petrology, v. 26, p. 356-362.

Brown, B. W., 1967, A Pliocene Tennessee River hypothesis for Mississippi: Southeastern Geology, v. 8, p. 81-84.

Cleaves, E. T., 1989, Appalachian Piedmont landscapes from the Permian to the Holocene: Geomorphology, v. 2, p. 159-179.

Conant, L. C., 1964, General remarks on the pre-Selma Cretaceous strata of western Alabama: U. S. Geological Survey Bulletin 1160-F, p. 97-101.

Costa, J. E., and Cleaves, E. T., 1984, The Piedmont landscape of Maryland: A new look at an old problem: Earth Surface Processes and Landforms, v. 9, p. 59-74.

Delcourt, P. A., 1980, Quaternary alluvial terraces of the Little Tennessee River Valley, east Tennessee, in Chapman, J., ed., The 1979 archaeological and geological investigation in the Tellico Reservoir: University of Tennessee Department of Anthropology Report 29, p. 110-121.

Goldstein, A., 1942, Sedimentary petrologic provinces of the northern Gulf of Mexico: Journal of Sedimentary Petrology, v. 12, p. 77-84.

Granger, D. E., Kirchner, J. W., and Finkel, R. C., 1997, Quaternary downcutting rate of the New River, Virginia, measured from differential decay of cosmogenic ^{26}Al and ^{10}Be in cave-deposited alluvium: Geology, v. 25, p. 107-110.

Granger, D. E., and Smith, A. L., 1998, Early Laurentide glaciation and creation of Ohio River dated by radioactive decay of cosmogenic Al-26 and Be-10 in proglacial sediments: Geological Society of America Abstracts with Programs, v. 30, no. 7, p. A298.

Grim, R. E., 1936, The Eocene sediments of Mississippi: Mississippi State Geological Survey Bulletin 30, 228 p.

Hayes, C. W., and Campbell, M. R., 1894, Geomorphology of the southern Appalachians: National Geographic Magazine, v. 6, p. 63-126.

Hayes, C. W., and Campbell, M. R., 1900, The relation of biology to physiography: Science, N.S. 12, no. 291, p. 131-133.

Holt, P. C., 1969, The distributional history of the biota of the southern Appalachians. Part I, invertebrates: Blacksburg, Virginia, Virginia Polytechnic Institute Research Division Monograph 1.

Houser, B., 1981, Erosional history of the New River, Southern Appalachians, Virginia: U.S. Geological Survey Open-file Report 81-771, 225 p.

Isphording, W. C., 1983, Interpretive mineralogy: Examples from the Miocene coastal plain sediments: Gulf Coast Association of Geological Societies, Transactions, v. 33, p. 295-306.

Johnson, D. W., 1905a, The distribution of fresh-water faunas as an evidence of drainage modification: Science, N.S., v. 21, p. 588-592.

Johnson, D. W., 1905b, The Tertiary history of the Tennessee River: Journal of Geology, v. 13, p. 194-231.

Johnson, D. W., 1939, Biologic evidence of capture: Journal of Geomorphology, v. 2, p. 88-91.

Johnson, D. W., 1941, Mussel distribution as evidence of drainage changes: Journal of Geomorphology, v. 4, p. 307-321.

Johnson, D. W., 1942, Mussel distribution as evidence of drainage changes: Journal of Geomorphology, v. 5, p. 59-72.

Kaye, J. M., 1974, Compositional sorting of topographically high Tennessee River gravels: a glacial hypothesis: Geology, v. 2, p. 45-47.

Long, S. H., 1875, Report of examinations and surveys with a view of improving the navigation of the Holston and Tennessee Rivers: Washington, D. C., Government Printing Office, 51 p.

Marcher, M. V., and Stearns, R. G., 1962, Tuscaloosa Formation in Tennessee: Geological Society of America Bulletin, v. 73, p. 1365-1386.

Mellen, 1939, Winston County mineral resources: Mississippi Geological Survey Bulletin 38, 169 p.

Milici, R. C., 1968, Mesozoic and Cenozoic physiographic development of the lower Tennessee River: in terms of the dynamic equilibrium concept: Journal of Geology, v. 76, p. 472-479.

Mills, H. H., 2000, Apparent increasing rates of stream incision in the eastern United States during the late Cenozoic: Geology, v. 28, p. 955-957.

Mills, H. H., and Wagner, J. R., 1985, Long-term change in the regime of the New River indicated by vertical variation in extent and weathering intensity of alluvium: Journal of Geology, v. 93, p. 131-142.

Monroe, W. H., Conant, L. C., and Eargle, D. H., 1946, Pre-Selma Upper Cretaceous stratigraphy of western Alabama: American Association of Petroleum Geologists Bulletin, v. 30, p. 187-212.

Parks, W. S., 1968, Geologic map of the Hebron Quadrangle, Tennessee: Tennessee Division of Geology, Geologic Map GM 440-NW, scale 1:24,000.

Parks, W. S., 1992, Four levels of terrace deposits and remnants of high level fluvial deposits in the Hatchie River Valley, Hebron area, Hardeman County, Tennessee: Mississippi Geology, v. 13, no. 4, p. 63-70.

Poag, C. W., and Sevon, W. D., 1989, A record of Appalachian denudation in postdrift Mesozoic and Cenozoic sedimentary deposits of the U.S. Middle Atlantic continental margin: Geomorphology, v. 2, p. 119-157.

Potter, P. E., 1955, The petrology and origin of the Lafayette gravel. Part 1. Mineralogy and petrology: Journal of Geology, v. 63, p. 1-38.

Reesman, A. L., and Godfrey, A. E., 1981, Development of the Central Basin of Tennessee by chemical denudation: Zeitschrift für Geomorphologie, v. 25, p. 437-456.

Reesman, A. L., and Stearns, R. G., 1989, The Nashville Dome - an isostatically induced erosional structure - and the Cumberland Plateau Dome - an isostatically suppressed extension of the Jessamine Dome: Southeastern Geology, v. 30, p. 147-174.

Rosen, N. C., 1969, Heavy minerals and size analysis of the Citronelle Formation of the Gulf Coastal Plain: Journal of Sedimentary Petrology, v. 39, p. 1552-1565.

Russell, E. E., 1964, Geologic map and mineral resource summary of the Pittsburg Landing Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 13-NE and MRS 13-NE.

Russell, E. E., 1967, Geologic map and mineral resource summary of the Michie Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 13-SW and MRS 13-SW.

Russell, E. E., 1968, Geologic map and mineral resource summary of the Counce Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 13-SE and MRS 13-SE.

Russell, E. E., and Wilson, C. W., 1970, Geologic map and mineral resource summary of the Thurman Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 23-NW and MRS 23-NW.

Russell, E. E., Wilson, C. W., and Jewell, J. W., 1972, Geologic map and mineral resource summary of the Pickwick Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 24-SW and MRS 24-SW.

Russell, E. E., and Parks, W. S., 1975, Stratigraphy of the outcropping Upper Cretaceous, Paleocene, and lower Eocene in western Tennessee (including descriptions of younger fluvial deposits): Tennessee Division of Geology Bulletin 75, 113 p.

Saucier, R. T., 1987, Geomorphological interpretations of late Quaternary terraces in western Tennessee and their regional tectonic implications: U. S. Geological Survey Professional Paper 1336-A, 19 p.

Self, R. P., 1984, Plio-Pleistocene drainage patterns and their influence on sedimentation in southeast Louisiana [Abs.]: Geological Society of America Abstracts with Programs, v. 16, p. 194.

Self, R. P., 1993, Late Tertiary to early Quaternary sedimentation in the Gulf Coastal Plain and lower Mississippi Valley: Southeastern Geology, v. 33, no. 2, p. 99-110.

Self, R. P., 2000, The pre-Pliocene course of the Lower Tennessee River as deduced from river terrace gravels in southwest Tennessee: Southeastern Geology, v. 39, in press.

Shaw, E. W., 1918, The Pliocene history of northern and central Mississippi: U.S. Geological Survey Professional Paper 108-H, p. 125-163. Simpson, C. T., 1900, The evidence of the Unionidae regarding the former courses of the Tennessee and other southern rivers: Science, N.S., v. 12, p. 133-136.

Smith, G. R., 1999, Using fish paleontology to reconstruct drainage histories: Geological Society of America Abstracts with Programs, v. 31, no. 7, p. A-443.

Swingle, G. D., 1959, Geology, mineral resources, and ground water of the Tennessee area, Tennessee: Tennessee Division of Geology Bulletin 61, 125 p.

Van der Schalie, H., 1939, Distributional studies of the Naiades as related to geomorphology: Journal of Geomorphology, v. 2, p. 251-257.

White, C. H., 1904, The Appalachian River versus a Tertiary trans-Appalachian river in eastern Tennessee: Journal of Geology, v. 12, p. 34-39.

Wilson, C. W., Russell, E. E., and Calvin, J. M. 1971, Geologic map and mineral resource summary of the Hookers Bend Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 23-SW and MRS 23-SW.

Wilson, C. W., Russell, E. E., and Jewell, J. W., 1982, Geologic map and mineral resource summary of the Savannah Quadrangle, Tennessee: State of Tennessee, Department of Conservation and Commerce, Division of Geology, GM 23-NW and MRS 23-NW.

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SALINITY RANGES OF LATE MISSISSIPPIAN INVERTEBRATES OF THE CENTRAL APPALACHIAN BASIN

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ABSTRACT

Late Mississippian (Chesterian) rocks from northern West Virginia and southwestern Pennsylvania contain numerous invertebrate fossils in the Wymps Gap Limestone Member of the Greenbrier Limestone, and in the Bickett Shale and Reynolds Limestone members of the Mauch Chunk Formation. The stratigraphic distribution of the various taxa was studied by multivariate analysis of abundance and taxon-count data sets obtained from bulk samples.

The primary control on the stratigraphic distribution of various taxa appears to have been salinity range during deposition. The patterns of faunal distribution, as well as the sedimentology of the rocks, allows separation of taxa into normal marine stenohaline and variable salinity euryhaline groups. Stenohaline taxa were mostly restricted to the Wymps Gap and Reynolds limestones and included all brachiopods (except for *Orthotetes* and *Lingula*), bryozoans, corals, trilobites, and echinoderms. The gastropods were divided between the stenohaline *Bullimorpha*, *Knightites*, and *Naticopsis* and the euryhaline *Bellerophon*, *Donaldina*, and *Straparollus*. Bivalves were mostly euryhaline, although *Sulcatopinna* and *Wilkingia* appear to have been stenohaline.

The Wymps Gap and Reynolds limestones record open marine deposition, whereas the Bickett Shale was apparently deposited in bays or estuaries, with restricted marine circulation, along a muddy shoreline during a period of regression.

INTRODUCTION

Late Mississippian (Chesterian) marine rocks of the north-central Appalachians are relatively fossiliferous, but few studies of their invertebrate paleoecology have been published (Simonsen, 1981, 1988; Rollins and Brezinski, 1988; Brezinski, 1988; Christopher, 1990). The present study was undertaken in an attempt to understand the environmental controls on the distribution of the relatively diverse and numerous invertebrates in the Greenbrier Limestone and lower part of the Mauch Chunk Formation exposed on the west limb of the Chesnut Ridge anticline in Monongalia County, West Virginia, and in adjacent Fayette County, Pennsylvania (Lake, 1998). The study interval is also known to contain a relatively high diversity of vertebrates. These include a variety of fish teeth and bone from the Greenbrier Limestone (Arkle, and others, 1979), as well as amphibians (Romer, 1969, 1970; Hotton, 1970), lungfish (Thomson, 1965), sharks, and acanthodians (Elliott and Taber, 1981) from the lower part of the Mauch Chunk Formation. Although fish teeth were found during the present study, vertebrates were not included in the analysis because of their relative scarcity.

The present study was conducted with bulk samples, but the relative abundance data were also converted to binary (presence/absence) and taxon-count data for comparison. An unexpected result from this was that both abundance and binary/taxon-count data yielded similar patterns from multivariate techniques indicating the underlying paleosalinity signal is strong and relatively insensitive to differences in relative

abundance between taxa.

GEOLOGIC SETTING

Stratigraphy

Chesterian rocks from northern West Virginia and southwestern Pennsylvania share many similarities. The stratigraphic nomenclature used in the present study (Figure 1) is based on previous studies of lithostratigraphic units from these two states. The following review of stratigraphy is presented as an aid in understanding prior usage of stratigraphic nomenclature in this region, and as an explanation for the stratigraphic nomenclature used in the present study.

Chesterian rocks in northern West Virginia are divided into the Greenbrier Limestone and Mauch Chunk Formation (Arkle and others, 1979). The Greenbrier Limestone in northern West Virginia has been subdivided into three members: 1) the basal Loyalhanna Limestone Member, a cross bedded sandy limestone; 2) an unnamed middle member of red and green shales and siltstones often correlated with the Taggard Formation of southern West Virginia; and 3) an upper member correlative with the Wymps Gap Limestone Member of the Mauch Chunk Formation of Pennsylvania (Flint, 1965; Berg and others, 1983; Brezinski, 1989a,c). The type section is at Wymps Gap on Chesnut Ridge, Fayette County, Pennsylvania (Hickok and Moyer, 1940), less than 2 km from the West Virginia border (UTM coordinates: Zone 17; 607,100 m E; 4,398,500 m N).

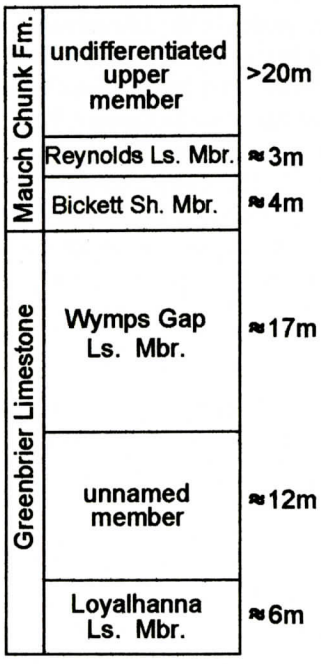

Figure 1. Stratigraphic column of Chesterian rocks in northern West Virginia and southwestern Pennsylvania. Approximate thicknesses are based on exposures at Greer Limestone Company quarry (Appendix 1). See text for discussion of individual rock units.

In the stratotype area of the Greenbrier Limestone in southern West Virginia, the Greenbrier is a group and is divided into several formations (Arkle and others, 1979). Various workers have applied Greenbrier Group stratigraphic nomenclature from rock sequences in southern West Virginia to those in northern West Virginia (e.g., Tilton, 1928; Busanus and Hoare, 1991). However, the Greenbrier Limestone consists of different facies on either side of the West Virginia Dome (Yielding and Dennison, 1986; Carney and Smosna, 1989; Smosna and Koehler, 1993). The Wymps Gap Limestone Member has often been lithologically correlated with the Union Limestone from southern West Virginia, however, the Union is typically oolitic (Kelleher and Smosna, 1993), whereas the Wymps Gap in northern West Virginia is a packstone/wackestone with interbedded shale layers.

The Lillydale Shale and Glenray Limestone members, of the Bluefield Formation of the Mauch Chunk Group in southern West Virginia, have also been lithologically correlated with units in northern West Virginia (Tilton, 1928; Hotton, 1970; Brezinski, 1989a,c; Busanus and Hoare, 1991). The Lillydale in southern West Virginia is a dark gray marine shale overlying the Alderson Limestone formation at the top of the Greenbrier Group and underlying the Glenray Limestone of the Bluefield Formation (Reger, 1926). The Glenray Limestone in Pocahontas County, at its furthest northern outcrop, is a laminated carbonate mudstone and is sparsely fossiliferous, unlike the Wymps Gap Limestone (personal observation, TWK). Based

on these lithologic differences between rocks in northern and southern West Virginia, we judge that rocks previously called Lillydale Shale and Glenray Limestone in northern West Virginia are simply the top couplet of alternating limestone and shale in the upper part of the Wymps Gap Limestone. As all the rocks in question are Chesterian in age (see below), biostratigraphic differentiation of the Lillydale and Glenray is not possible.

The Greenbrier Limestone and Mauch Chunk formations of northern West Virginia are separated at the top of the Wymps Gap Limestone Member (Figure 1). The Mauch Chunk predominantly consists of clastic rocks, with the notable exception of the Reynolds Limestone Member. Mauch Chunk rocks between the Wymps Gap and Reynolds limestones consist of red, green, and gray shales, siltstones, and sandstones. These rocks are similar to the Bickett Shale Member of the Bluefield Formation, which underlies the Reynolds in southern West Virginia (Reger, 1926). Because the Reynolds conformably overlies these rocks in northern West Virginia, and they appear to be from similar facies in both north and south, it seems reasonable to extend the name Bickett Shale from southern West Virginia as was done by other workers (Tilton, 1928; Hotton, 1970; Brezinski, 1989a,c; Busanus and Hoare, 1991). The Reynolds Limestone represents the last marine transgression in the Chesterian of this region (Brezinski, 1989a,b,c). The Reynolds is an easily recognizable unit that occurs throughout the subsurface of West Virginia as the "Little Lime" of oil and gas drillers, and is readily correlated from its type area (Reger, 1926) in southern West Virginia. The remaining Mauch Chunk above the Reynolds in northern West Virginia is not divided into members. These rocks are apparently nonmarine and are similar to the red-bed-dominated Hinton Formation of southern West Virginia.

Age

Rocks of the Greenbrier Limestone and Mauch Chunk Formation in northern West Virginia are Chesterian in age, based on previous biostratigraphic studies. The Loyalhanna Limestone Member contains the brachiopods *Anthracospirifer pellaensis, Composita subquadrata,* and *Orthotetes kaskaskiensis* (Brezinski, 1989b). This assemblage is most likely age-equivalent to the Ste. Genevieve Limestone (Carter and Carter, 1970) of the Illinois basin, which is early Chesterian (Maples and Waters, 1987). The Wymps Gap Limestone Member is middle Chesterian (pre-Glen Dean Limestone) based on conodonts (Horowitz and Rexroad, 1972). The Reynolds Limestone Member is correlative with the middle Chesterian Glen Dean Limestone of the Illinois basin on the basis of conodonts (Rexroad and Clark, 1960) and total fauna (Henry and Gordon, 1992). There are no late Chesterian marine rocks in northern West Virginia, although marine rocks of this age outcrop in southern West Virginia (Henry and Gordon, 1992).

Lithologies and Depositional Environments

The Wymps Gap Limestone typically consists of gray, interbedded fossiliferous wackestone, packstone, and shale. Carney and Smosna (1989), in a detailed study of carbonate facies of the Greenbrier Limestone, characterized the Wymps Gap Limestone Member as an open marine facies on the basis of the highly diverse fauna, including stenohaline (normal marine salinity, 35 parts per thousand) taxa such as pelmatozoans (crinoids and blastoids), brachiopods, and bryozoans. They noted evidence for low energy conditions, such as lack of abrasion of skeletal particles and relatively high carbonate mud content. They also noted evidence for deposition in the photic zone consisting of calcareous algae and oncolites. Brezinski (1989a) noted the absence of current indicators in the Wymps Gap Limestone. This, along with evidence of a contemporaneous carbonate sand shoal to the northeast in Pennsylvania, led him to conclude that the Wymps Gap was deposited below average storm wave base at 40 to 50 m water depth.

The Bickett Shale typically consists of green, red, tan, and brown silty shale, clay shale, and

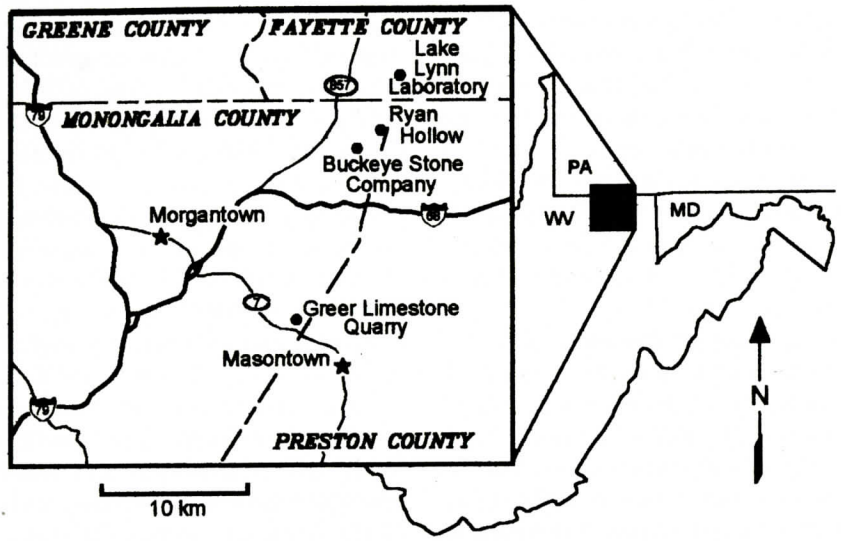

Figure 2. Map of northern West Virginia and southwestern Pennsylvania showing localities studied. Detailed locality information listed in Appendix 1.

minor sandstone in small paleo-channels. Paleosols also are present as indicated by green-colored root mottles and plant debris in red sediments. Brezinski (1989a) interpreted the Bickett Shale as representing prograding shoreline and alluvial plain deposits.

The Reynolds Limestone typically is a brownish gray, fossiliferous wackestone and packstone. The Reynolds Limestone was deposited during a transgression and represents the last Mississippian marine sediments deposited in the study area (Brezinski, 1989a). Younger marine rocks are present in the Chesterian of southern West Virginia (e.g., Henry and Gordon, 1992).

Paleogeographic reconstructions of depositional environments during Greenbrier Limestone and Mauch Chunk Formation deposition are presented in Carney and Smosna (1989) and Brezinski (1989a, 1999b).

FAUNAS

Late Mississippian faunas of the Appalachian basin have been documented in various studies extending back more than one hundred years (e.g., Meek, 1871; Girty, 1923; Butts, 1940-1941). More recent studies employing modern taxonomy include Hoskins and others (1983) on various invertebrates; Brezinski (1988, 1999a) on trilobites; Brezinski (1989b) on brachiopods, molluscs, and bryozoans; Busanus and Hoare (1991) and Hoare (1993) on bivalves; and Henry and Gordon (1992) on brachiopods. Thien and Nitecki (1974) described gastropods from the Chesterian of the Illinois Basin, many of which also occur in the Appalachian basin. The illustrations in these various references are very useful for identification of the genera recognized in this study. Most of the genera are also treated in the appropriate volumes of the *Treatise on Invertebrate Paleontology*. The reader should consult these various references for illustrations of the Chesterian taxa treated in the present study.

DATA AND METHODS

Outcrops

Four outcrops were sampled in detail (Figures 2 and 3, Appendix 1). Three are in Monongalia County, West Virginia: 1) Greer Limestone Company quarry, 2) Buckeye Stone Company quarry, and 3) Ryan Hollow. The fourth is in Fayette County, Pennsylvania: 4) Lake Lynn Laboratory quarry.

The stratigraphic sequence exposed in the

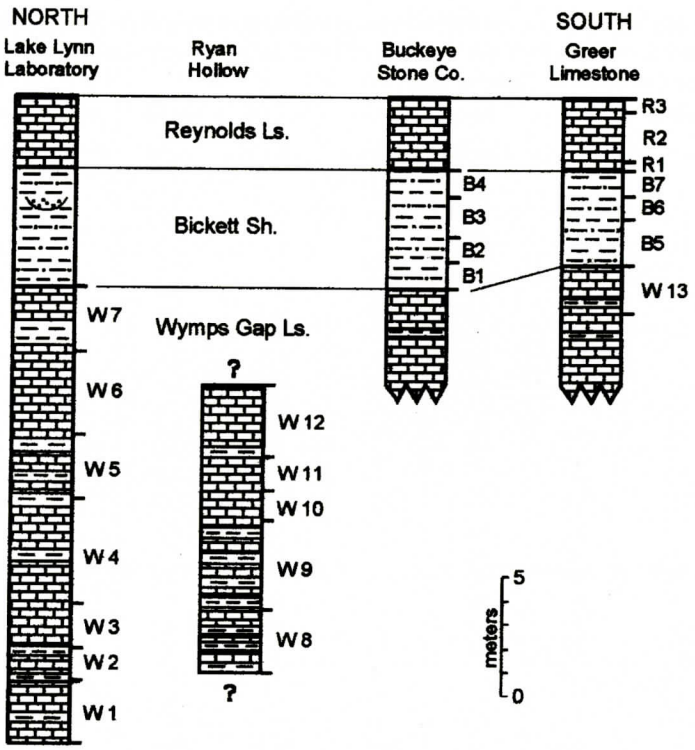

Figure 3. Stratigraphic columns for study localities (Figure 2 and Appendix 1) showing sample locations and generalized lithologies. No horizontal scale. Note the fluvial paleo-channel in the Bickett Shale at Lake Lynn Laboratory. The placement of the upper and lower contacts of the Wymps Gap Limestone at Ryan Hollow are uncertain.

Greer Limestone Company quarry ranges from the Loyalhanna through the upper Mauch Chunk (Figure 1), but only the upper part of the Wymps Gap Limestone, Bickett Shale, and Reynolds Limestone were sampled (Figure 3). This section was originally described by Tilton (1928), who provided a reasonably accurate section description, but applied stratigraphic names from southern West Virginia, as discussed above in the stratigraphy section. Lucke (1939, p. 43) also provided a measured section. Hotton (1970, pl. 2) illustrated the Bickett Shale and Reynolds Limestone exposure. Busanus and Hoare (1991) provided a measured section following Tilton's (1928) stratigraphic nomenclature.

The Buckeye Stone Company quarry also contains the interval from the Loyalhanna through the upper Mauch Chunk. It is an active quarry with the rocks exposed in nearly vertical highwalls. Only the Bickett Shale could be safely sampled at this locality (Figure 3).

The Ryan Hollow outcrop is an abandoned quarry from the late nineteenth or early twentieth century. Only the Wymps Gap Limestone Member is exposed and the weathered outcrop yields abundant fossils. The exact stratigraphic position of the outcrop within the Wymps Gap Limestone is unknown as there is no evidence of the lower or upper contacts (Figure 3).

The Lake Lynn Laboratory quarry contains the interval from the Wymps Gap Limestone to the upper Mauch Chunk. Because of the nearly vertical highwall produced by quarrying, only the Wymps Gap could be sampled. The Bickett Shale in the highwall shows typical green and red lithologies, as well as a distinctive sandstone channel of probable fluvial origin (Figure 3).

Table 1. Percent abundance data for individual taxa arranged by guilds. Sample abbreviations: W = Wymps Gap Ls., B = Bickett Sh., R = Reynolds Ls. Guild abbreviations, and abbreviated taxa names, are defined in Table 3. Sample size is n.

Guild	T-Brach	T-Brach	T-Brach	T-Brach	T-Brach	P-Brach	P-Brach	P-Brach	Bryozo-ans	Corals	
Sam-ples	Anthraco	Cleiothy	Composita	Eumetria	Orthotetes	Diaphrag	Ovatia	Productids	Bryozo-ans	Rugosa	
W1	22.4	1.8	12.1	0.0	17.5	10.8	1.4	14.4	7.2	0.5	
W2	3.2	1.6	9.7	0.0	14.5	12.9	1.6	8.1	9.7	0.0	
W3	20.8	10.4	10.7	0.3	18.5	5.4	1.0	6.4	12.1	1.0	
W4	30.9	0.0	8.1	0.0	25.3	7.0	3.2	6.3	6.0	0.7	
W5	23.5	0.0	17.7	0.0	5.9	23.5	0.0	0.0	14.7	2.9	
W6	18.3	2.7	12.3	0.3	9.3	13.6	3.0	16.3	4.3	4.0	
W7	36.4	0.0	13.0	0.0	8.4	11.0	2.6	11.0	5.8	0.0	
W8	3.3	5.0	10.0	0.0	0.0	26.7	8.3	11.7	3.3	13.3	
W9	2.7	4.0	12.4	0.3	0.7	33.9	5.7	13.1	11.1	5.0	
W10	3.5	6.9	20.7	0.0	0.0	24.1	3.5	6.9	6.9	0.0	
W11	17.0	0.0	17.0	0.0	34.0	5.7	0.0	1.9	22.6	0.0	
W12	15.8	0.0	5.3	0.0	30.7	0.0	0.0	0.0	48.3	0.0	
W13	31.3	0.0	37.5	0.0	18.8	0.0	0.0	0.0	6.3	0.0	
B1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	
B2	0.0	0.0	0.0	0.0	2.4	0.0	0.0	0.0	0.0	0.0	
B3	0.0	0.0	0.0	0.0	2.9	0.0	0.0	0.0	13.3	0.0	
B4	0.0	0.0	0.0	0.0	2.0	0.0	0.0	0.0	0.0	0.0	
B5	0.0	0.0	0.0	0.0	2.6	0.0	0.0	0.0	0.0	0.0	
B6	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	
B7	0.0	0.0	0.0	0.0	0.6	0.0	0.0	0.0	0.0	0.0	
R1	11.1	0.0	12.3	0.0	3.0	14.5	2.1	0.0	1.7	0.0	
R2	32.5	0.0	19.0	0.0	20.2	2.4	0.2	0.0	12.4	0.0	
R3	16.7	0.0	16.7	0.0	0.0	16.7	6.7	0.0	0.0	0.0	
Guild	Tri	Tri	Gastro	Gastro	Gastro	Gastro	Gastro	Gastro	Epi-Biv	Epi-Biv	
Sam-ples	Kaskia	Paladin	Bullimo	Donaldina	Naticopsis	Bellero	Knight	Straparollus	Aviculo	Limipect	
W1	0.0	2.2	1.4	0.0	0.0	0.5	0.0	0.0	1.4	0.0	
W2	0.0	1.6	4.8	0.0	0.0	8.1	0.0	0.0	0.0	0.0	
W3	0.0	5.0	0.0	0.0	0.0	2.0	0.0	0.0	0.7	0.0	
W4	0.4	2.8	0.4	0.0	0.0	1.4	0.4	0.0	1.4	0.0	
W5	0.0	0.0	2.9	0.0	0.0	5.9	0.0	0.0	0.0	0.0	
W6	1.3	1.3	0.7	0.0	0.0	2.3	1.3	1.0	0.7	0.0	
W7	0.7	2.0	0.0	0.0	0.0	2.6	0.7	0.0	1.3	0.0	
W8	3.3	3.3	0.0	0.0	0.0	0.0	3.3	1.7	0.0	0.0	
W9	0.0	1.3	1.3	0.3	0.3	3.7	0.3	0.7	0.7	0.0	
W10	0.0	10.3	0.0	0.0	0.0	3.5	0.0	13.8	0.0	0.0	
W11	0.0	0.0	0.0	1.9	0.0	0.0	0.0	0.0	0.0	0.0	
W12	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	
W13	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	
B1	0.0	0.0	0.0	1.2	0.0	6.1	0.0	2.4	22.0	0.0	
B2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	23.8	12	
B3	0.0	0.0	0.0	0.0	1.0	26.7	0.0	1.0	10.5	11.4	
B4	0.0	0.0	0.0	4.0	0.0	5.3	0.7	0.0	14.0	5.3	
B5	0.0	0.0	0.0	2.6	0.0	2.6	0.0	0.0	7.9	0.0	
B6	0.0	0.0	0.0	9.2	0.0	3.1	0.0	0.0	16.9	0.0	
B7	0.0	0.0	0.0	6.7	0.0	0.0	0.0	0.0	2.8	2.8	
R1	0.0	1.3	0.0	1.3	0.4	39.2	3.8	0.0	0.0	0.0	
R2	0.0	0.2	0.0	0.0	2.4	6.8	0.0	0.0	1.6	0.0	
R3	0.0	3.3	0.0	0.0	0.0	16.7	0.0	0.0	0.0	0.0	
Guild	Epi-Biv	Epi-Biv	In-Biv	In-Biv	In-Biv	In-Biv	In-Biv	In-Biv	In-Biv	0.0	
Sam-ples	Streblo	Sul-catopinna	Ectogram	Edmondia	Paleyoldia	Phestia	Sanguin	Schizodus	Sphe-notus	Wilkin-gia	n
W1	0.0	0.0	0.0	1.8	0.0	0.0	0.9	0.0	1.4	2.7	223
W2	0.0	0.0	1.6	4.8	0.0	0.0	16.1	0.0	0.0	1.6	62
W3	0.0	0.0	0.0	0.7	0.0	0.0	1.3	1.3	1.7	0.7	298
W4	0.0	0.0	0.7	2.1	0.0	0.0	0.4	0.4	1.8	0.7	285
W5	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.9	0.0	34
W6	0.0	0.0	0.0	1.7	0.0	0.0	0.7	0.7	0.3	4.0	301
W7	0.0	0.0	2.0	1.3	0.0	0.0	0.7	0.0	0.0	0.7	154

Table 1. Percent abundance data for individual taxa arranged by guilds. Sample abbreviations: W = Wymps Gap Ls., B = Bickett Sh., R = Reynolds Ls. Guild abbreviations, and abbreviated taxa names, are defined in Table 3. Sample size is n.

W8	0.0	0.0	0.0	0.0	0.0	0.0	3.3	0.0	0.0	3.3	60
W9	0.0	0.7	0.0	0.0	0.0	0.0	0.3	0.0	0.7	0.7	298
W10	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	29
W11	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	53
W12	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	114
W13	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	6.3	16
B1	4.9	0.0	13.4	3.7	0.0	11.0	8.5	25.6	1.2	0.0	82
B2	0.0	0.0	10.7	13.1	0.0	14.3	1.2	33.3	0.0	0.0	84
B3	0.0	1.0	5.7	9.5	0.0	1.9	7.6	4.8	2.9	0.0	105
B4	4.0	0.0	5.3	8.7	0.0	18.0	10.7	18.0	4.0	0.0	150
B5	0.0	0.0	0.0	13.2	13.2	23.7	0.0	34.2	0.0	0.0	38
B6	0.0	0.0	0.0	9.2	10.8	13.9	7.7	27.7	1.5	0.0	65
B7	0.0	0.0	2.2	29.6	1.1	15.6	15.1	21.8	1.7	0.0	179
R1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	9.4	235
R2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.2	499
R3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	3.3	20.0	30
											3394

Table 2. Means and standard deviations of taxa and specimen counts (n) from 23 bulk samples listed in Table 1.

Bulk samples	Taxa, mean	Taxa, st. dev.	Specimens (n), mean	Specimens (n), st. dev.
Wymps Gap Limestone (13)	13.8	6.2	148	116
Bickett Shale (7)	10.6	2.4	100	49
Reynolds Limestone (3)	10.3	2.1	255	235

Table 3. Composition of guilds and primary references, other than the *Treatise on Invertebrate Paleontology*, for identification of taxa [bracketed taxa were too rare to include in Table 1].

Tethered Brachiopods (T-Brach): *Anthracospirifer, Cleiothyridina, Composita, Eumetria, [Lingula], Orthotetes* (Henry and Gordon, 1992)

Productid Brachiopods (P-Brach): *Diaphragmus, Ovatia*, productids sp. (Henry and Gordon, 1992)

Bryozoans: individual fenestrate and rhomboporid colonies, unidentified to genus level

Corals: solitary rugose corals, unidentified to genus level

Trilobites (Tri): *Paladin, Kaskia* (Brezinski, 1988, 1999)

Gastropods (Gastro): *Bellerophon, Bullimorpha, Donaldina, [Eotrochus], Knightites, Naticopsis, [Platyzona], Straparollus* (Thein and Nitecki, 1974)

Epifaunal bivalves (Epi-Biv): *Aviculopecten, Limipecten, Streblochondria, Sulcatopinna* (Busanus and Hoare, 1991; Hoare, 1993)

Infaunal bivalves (In-Biv): *Ectogrammysia, Edmondia, Paleyoldia, Phestia, [Prothyris], Sanguinolites, Schizodus, Sphenotus, Wilkingia* (Busanus and Hoare, 1991; Hoare, 1993)

[Echinoderms: crinoid and blastoid columnals]

Samples

Approximately 3400 fossil specimens were isolated from a total of 103 bulk samples collected at the four localities. Samples were approximately 10 kg in size, although sample size was largely controlled by ease of sampling (e.g. well-indurated versus poorly-indurated limestone, or limestone versus shale). Limestone samples consisted of either friable rock or slabs. Individual fossils were recovered from friable rock. Slabs had all specimens on their exposed surfaces identified and counted. Bulk samples of shales were mechanically disaggregated and both fossils in the round and molds were counted. Individual valves of bivalves were counted and divided by two so as to not overestimate their relative abundance. Brachiopods were counted by pedicle valves.

It turned out that most bulk samples had sparse data, so lithologically similar samples from adjacent stratigraphic levels, within the same stratigraphic member, were combined into a total of 23 samples (Table 1, Figure 3). This substantially improved the data density, both in number of taxa and abundance of specimens, for the average sample (Table 2). Dennison and Hay (1967) suggested a sample size of 300 specimens was sufficient for determination of relative abundance of taxa. The average sample size in this study ranged from a low of 100 specimens for the Bickett Shale to a high of 255 for the Reynolds Limestone (Table 2), with individual samples ranging from 16 to 499 specimens. Practical restrictions on both sample size (rock volume) and time prevented larger sample sizes from being collected and analyzed.

Fossil Taxa

Specimens recovered in the present study were identified primarily to the genus level, except for bryozoans and corals (Tables 1 and 3). Genus-level identifications were determined to be sufficient because the often poor or fragmentary preservation of specimens created uncertainty in species-level identifications for all but the best preserved specimens from the bulk samples. Additionally, it is reasonable to assume that the genus level is sufficient for paleoecologic analysis as the ecologic differences between species within a single genus were probably much less than that between genera.

A total of 30 taxa were recognized among the 23 samples. Additional taxa found in the samples, *Lingula, Eotrochus, Platyzona, Strophostylus*, and *Prothyris* (Table 3), occurred only once in the bulk samples and were not included in Table 1. Crinoid and blastoid columnals (calyx plates of both crinoids and blastoids were found only in outcrop surface collections) were relatively common in both the Wymps Gap and Reynolds limestones, but were not included in Table 1, or the statistical analyses, as there was no way to determine the number of individuals represented by disarticulated columnals. They are mentioned here in order to acknowledge their presence as stenohaline taxa (Dodd and Stanton, 1990).

Guilds

To test for large-scale paleoecologic patterns involving groups of taxa, individual taxa were grouped into *a priori* defined guilds prior to multivariate analysis (Table 3). The guild concept in paleoecology categorizes organisms by food source, morphology, and life habits (Bambach, 1983; Dodd and Stanton, 1990). Bambach (1983, p. 728) extended the guild concept beyond that used in ecology to erect groupings based not only on similar exploitation of environmental resources, but also on similarities of space utilization and taxonomic membership. Droser and others (1997, 2000) used the term Bambachian megaguilds to denote such groupings based on gross taxonomic composition and adaptive strategies, which is how our guilds are defined. Grouping taxa into guilds reduced the number of variables from 30 to nine, which substantially reduces the dimensionality of the data set so that patterns within the data are more easily recognized (Gauch, 1982).

The guilds were defined as follows (Table 3). Brachiopods were divided into tethered (epifaunal attached-suspension-feeding guild of Bambach, 1983) and productids (epifaunal reclining-suspension-feeding guild of Bam-

bach, 1983) to see if position relative to the substrate was important in their distribution. Tethered brachiopods included those with a functional pedicle and consisted almost entirely of articulate brachiopods, except for the inarticulate *Lingula*, which was rare but occurred in the Bickett Shale. Tethered brachiopods were suspension feeders that held their commissure at a high angle to the substrate and were anchored by a pedicle (Rowell and Grant, 1987). Productid brachiopods have a strongly convex pedicle valve and a flat or concave brachial valve. These suspension feeders were free-lying, sitting on their pedicle valves with their commissure parallel to the substrate (Rowell and Grant, 1987). Bryozoans also are suspension feeders and included both fenestrate and rhomboporid colonies of the epifaunal attached-erect-suspension feeding guild (Bambach, 1983). For corals, only solitary rugose corals were counted, although fragments of colonial rugose and tabulate corals were sometimes observed. Corals are suspension-feeding passive carnivores that capture prey with their stinging cells (cnidoblasts) (Oliver and Coates, 1987). Trilobites were presumably mobile benthic detritus feeders (Robison, 1987) of the infaunal shallow-active-deposit-feeding guild (Bambach, 1983). Gastropods were also mobile organisms, but the feeding behavior of Paleozoic archaeogastropods is poorly understood. Most archaeogastropods today are grazing herbivores, but a few are scavengers or sluggish carnivores (Peel, 1987). Epifaunal bivalves were all suspension feeders (Pojeta, 1987) within the epifaunal attached-low-suspension-feeding guild (Bambach, 1983). Infaunal bivalves included both suspension and detritus feeders of the infaunal shallow-active-feeding guild (Bambach, 1983). Modern nuculoids are detritus feeders (Pojeta, 1987), so the nuculoids *Paleyoldia* and *Phestia* were presumably detritus feeders within the infaunal bivalve guild.

DATA SETS

Abundance data for guilds were derived by summing the data for taxa in each guild. Taxon-count data were constructed by converting abundance data to binary (presence/absence) data and then tabulating the number of taxa in each guild. All the analyzed data sets were derived from Table 1, which lists taxa by percent abundance (available in Microsoft Excel® format from TWK). There was no practical way to standardize sample size, so data are reported as percent abundance. Original raw data counts can be reconstructed by multiplying the percent data by sample size (n) and dividing by 100.

Abundance data are collected through bulk sampling and are more difficult to obtain because each sample must be carefully disaggregated and all specimens identified and counted. There is also the problem of sample size standardization. Should it be done by standardized rock sample mass, standardized counts (e.g., 300), or percent abundance? Standardization of rock sample mass is problematic because of unknown and, presumably, variable rates of sedimentation. Also, different lithologies yield different types of data. Shales readily disaggregate to yield isolated specimens, whereas well-cemented limestones must be studied by counting specimens on bedding surfaces. Standardized counts are difficult to obtain for low abundance samples. Percent abundance can be obtained from whatever size sample is available. However, there is the potential for spurious negative correlations created by closed data sets where all values sum to the same amount, e.g., 100 (Smith, 1999). This potential problem can be evaluated by comparing the matrices of correlation coefficients (r values) of the guilds for percent abundance versus raw abundance data. For both data sets the epifaunal and infaunal bivalves were negatively correlated with the tethered brachiopods, productid brachiopods, bryozoans, corals, and trilobites. These negative correlations were somewhat stronger for the percent abundance data than the raw abundance data, but the patterns of negative correlation were essentially the same. The only statistically significant ($p<0.05$) negative correlation produced by the percent abundance data, but absent from the raw abundance data, was between gastropods and tethered brachiopods ($r = -0.45$). Another test for spurious negative correlations is to calculate the correlation coefficient be-

tween the percent abundance and raw abundance data for the guilds. This yielded r = 0.71 ($p<0.0001$), which also is essentially the same value, r = 0.73, as the correlation between the percent abundance and taxon-count data for the guilds. These high r values suggest that any spurious negative correlations created by the closed data of percent abundance have not substantially altered the underlying statistical pattern of faunal distribution in the rocks.

Multivariate Analysis

Various multivariate techniques were used to explore for patterns in the data sets derived from Table 1. These included correlation matrices, binary plots, Q- and R-mode cluster analysis, Q- and R-mode multidimensional scaling, principal components analysis (PCA), factor analysis, and discriminant analysis. All techniques yielded similar results, but those from factor analysis were judged best for presentation because this technique simultaneously evaluates the correlations between variables and plots (ordinates) the relationships between all samples, or between all variables.

Factor analysis is a commonly used multivariate technique (e.g., Swan and Sandilands, 1995) that creates linear combinations of the original variables in order to better understand the underlying "factors" controlling the variation within the data. The statistical software package MINITAB for Windows Release 12 (Minitab, Inc., 1998) was used for the analysis. The analysis begins with a principal components analysis of the correlation matrix of the variables. This determines the number of principal components with eigenvalues close to or greater than 1.0, which are those that explain as much or more variance than a single original variable. Factor analysis is then performed where the number of factors retained equals the number of principal components with eigenvalues close to or greater than 1.0. The goal of factor analysis is to construct new variables, or factors, that are linear combinations of the original variables. A successful factor analysis will substantially reduce the number of variables to be considered by combining original variables into a few factors that represent the essential underlying structure in the data. The investigator must interpret what these factors represent by comparing them with the original data. The correlations between the original variables and the factors are termed loadings. Also, each factor is uncorrelated with other factors. The factors were then orthogonally rotated by the varimax criterion. Varimax rotation attempts to move each factor in multivariate space to a position such that the projection of the original variables are either near the extremities or near the origin of the factors (Davis, 1986). Thus, each of the original variables will be either strongly correlated (positive or negative) or uncorrelated with each of the rotated factors, making interpretation easier. Because the rotation is orthogonal, the rotated factors remain uncorrelated with one another. Communalities are an estimate of how much of the original variance for each variable is explained by the rotated components. Communalities are calculated by summing the squared correlations (loadings) of each variable with the retained factors.

Factor analysis was performed on the guild data sets, including percent abundance data (Table 4), raw abundance data (Table 5), and binary/taxon-count data (Table 6). For both the percent and raw abundance data, three factors were extracted and a total of 80 percent and 74 percent of the variance, respectively, were explained by factor analysis. Only two factors were extracted from the binary/taxon-count data set, which explained 75 percent of the variance.

Besides presentation of the loadings from factor analysis (Tables 4-6), the results of factor analysis can also be presented by ordination. Ordination is a technique for arranging samples or taxa in a low-dimensional space so that similar objects are close together. Ordination is a useful technique in both ecology and paleoecology for detecting environmental gradients such as moisture gradients associated with elevation and its control on plant communities (Gauch, 1982), or paleo-oxygen gradients controlling benthic community composition (Etter, 1999). Factor analysis was used to ordinate both samples (Q-mode) and variables (R-mode). For Q-

Table 4. Sorted rotated factor loadings and communalities for percent abundance data. Whether a loading is positive or negative with reference to a given factor is arbitrary, but the difference in sign is significant when comparing loadings within a given factor.

Guilds	Factor 1	Factor 2	Factor 3	Communality
T-Brachs	-0.893	-0.025	-0.237	0.854
In-Biv	0.857	0.436	-0.061	0.928
Epi-Biv	0.757	0.468	0.096	0.802
Bryozoans	-0.711	0.232	-0.100	0.569
P-Brachs	-0.173	-0.920	0.038	0.878
Corals	0.133	-0.808	-0.321	0.774
Trilobites	-0.161	-0.766	0.276	0.689
Gastropods	0.238	-0.015	0.931	0.923
Variance	2.740	2.551	1.126	6.417
% Variance	34.3	31.9	14.1	80.2

Table 5. Sorted rotated factor loading and communalites for raw abundance data. Whether a loading is positive or negative with reference to a given factor is arbitrary, but the difference in sign is significant when comparing loadings within a given factor.

Guilds	Factor 1	Factor 2	Factor 3	Communality
Corals	0.944	0.028	-0.068	0.897
P-Brachs	0.937	-0.169	-0.106	0.918
Trilobites	0.611	-0.340	-0.255	0.554
T-Brachs	0.175	-0.893	-0.202	0.868
Bryozoans	0.115	-0.807	-0.265	0.735
Gastropods	0.045	-0.545	0.311	0.396
In-Biv	-0.119	0.150	0.861	0.778
Epi-Biv	-0.173	0.031	0.853	0.758
Variance	2.233	1.914	1.757	5.904
% Variance	27.9	23.9	22.0	73.8

Table 6. Sorted rotated factor loadings and communalities for binary/taxon-count data. Whether a loading is positive or negative with reference to a given factor is arbitrary, but the difference in sign is significant when comparing loadings within a given factor.

Guilds	Factor 1	Factor 2	Communality
P-Brachs	0.899	-0.217	0.855
Trilobites	0.845	-0.036	0.715
T-Brachs	0.836	-0.375	0.839
Corals	0.780	0.132	0.626
Bryozoans	0.694	-0.484	0.716
Gastropods	0.567	0.518	0.590
In-Biv	-0.077	0.897	0.810
Epi-Biv	-0.187	0.890	0.827
Variance	3.673	2.305	5.978
% Variance	45.9	28.8	74.7

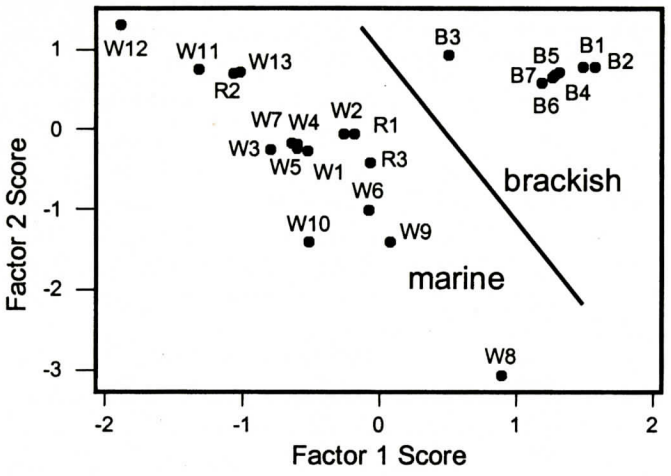

Figure 4. Q-mode ordination of percent abundance data using scores from factor analysis of guilds. Note the clear separation of the Bickett Shale samples (B) from the Wymps Gap (W) and Reynolds (R) limestones. Factors 1 and 2 combined explain 66 percent of the variance.

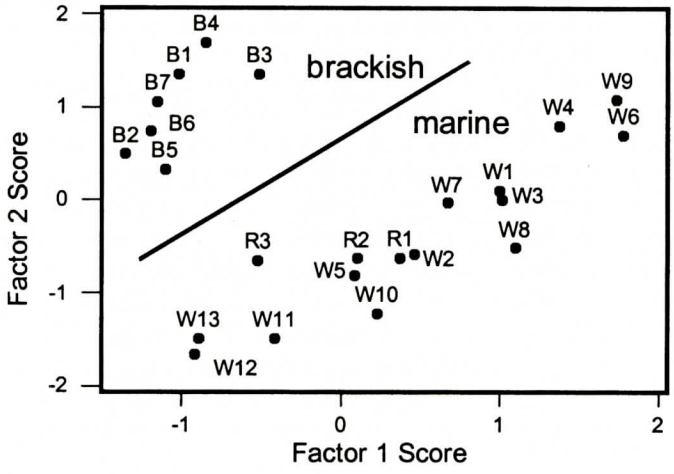

Figure 5. Q-mode ordination of binary/taxon-count data using scores from factor analysis of guilds. Note the clear separation of the Bickett Shale samples (B) from the Wymps Gap (W) and Reynolds (R) limestones. Factors 1 and 2 combined explain 75 percent of the variance.

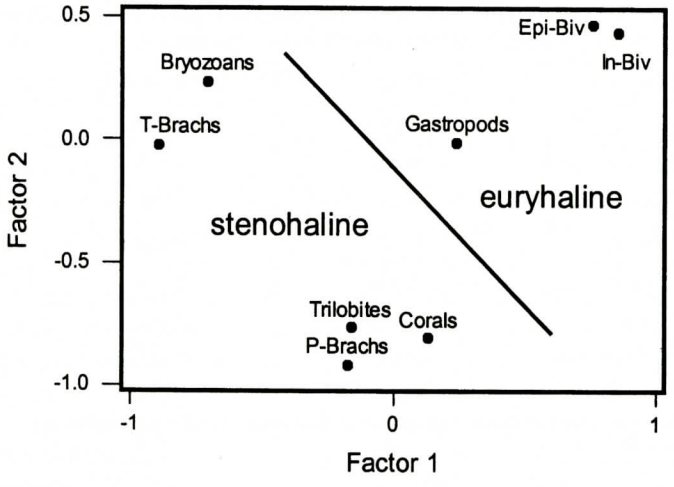

Figure 6. R-mode ordination of percent abundance data using loadings from factor analysis of guilds. This is a graphic representation of the data in Table 4. The guilds in the euryhaline field contain both euryhaline and stenohaline taxa (Table 7).

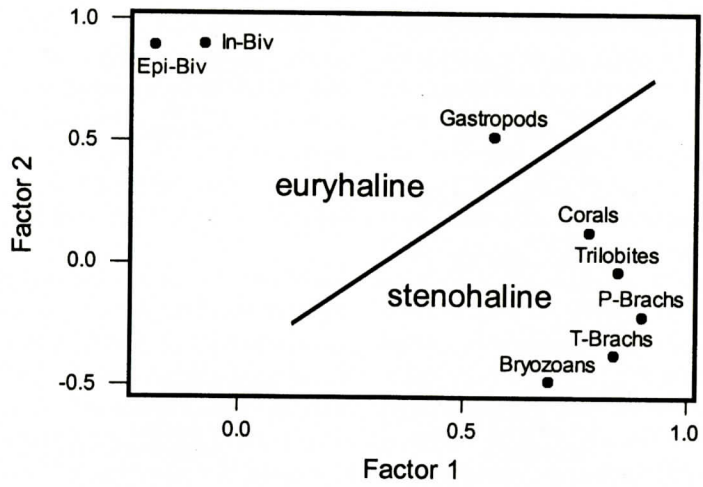

Figure 7. R-mode ordination of binary/taxon-count data using loadings from factor analysis of guilds. This is a graphic representation of the data in Table 6. The guilds in the euryhaline field contain both euryhaline and stenohaline taxa (Table 7).

mode ordination the values of the original variables are used to calculate a score for each sample on each factor. Samples that have high values for a given variable(s) will have high scores on those factors highly correlated with the variable(s), and vice versa. Q-mode ordination plots were produced for both percent abundance data (Figure 4) and binary/taxon-count data (Figure 5). These plots show the similarities/dissimilarities between samples. For R-mode ordination, the loadings of the original variables on the factors are plotted. This produces a map of similarities/dissimilarities between the variables. Once again both percent abundance data (Figure 6) and binary/taxon-count data (Figure 7) were plotted. The Q- and R-mode ordination plots of the raw abundance data were similar to the percent abundance data ordinations and, thus, were redundant and are not presented.

RESULTS AND DISCUSSION

The results of multivariate analysis indicates clear patterns are present in both the abundance data and binary/taxon-count data arranged by guilds (Tables 4-6, Figures 4-7). Q-mode ordination clearly separates the Bickett Shale samples from the Wymps Gap and Reynolds limestone samples (Figures 4 and 5). Also, there is no separation of the Wymps Gap from the Reynolds samples. R-mode ordination separates the tethered and productid brachiopods, bryozoans, corals, and trilobites from the infaunal and epifaunal bivalves, with the gastropods plotting between these two groupings (Figures 6 and 7). We interpret these patterns to be a reflection of paleosalinity differences between the open marine Wymps Gap and Reynolds limestones, and the brackish water Bickett Shale.

Living aquatic invertebrates can be categorized by their salinity ranges. Organisms restricted to normal marine conditions (35 parts per thousand salinity) are termed stenohaline (Allaby, 1985; Dodd and Stanton, 1990). These organisms are unable to regulate ionic concentrations within their cells and are killed by osmotic pressure changes under conditions of variable salinity (Pearse and Gunter, 1957; Croghan, 1983). Organisms capable of regulating ionic concentrations to some degree can tolerate variable salinity and are termed euryhaline (Allaby, 1985). Thus, euryhaline organisms may be found in a variety of environments including normal marine, brackish (5-35 parts per thousand salinity), and hypersaline (>35 parts per thousand salinity). Studies of modern and ancient bryozoans, brachiopods, corals, and echinoderms indicate these groups are mostly stenohaline with few exceptions, whereas mol-

Table 7. Interpreted salinity ranges of taxa.

Stenohaline (narrow)	Euryhaline (wide)
Tethered Brachiopods	Tethered Brachiopods
Anthracospirifer	*Lingula*
Cleiothyridina	*Orthotetes*
Composita	
Eumetria	
Productid Brachiopods	
Diaphragmus	
Ovatia	
Productids sp.	
Bryozoans	
Corals	
Trilobites	
Kaskia	
Paladin	
Gastropods	Gastropods
Bullimorpha	*Bellerophon*
Knightites	*Donaldina*
Naticopsis	*Straparollus*
Epifaunal Bivalves	Epifaunal Bivalves
Sulcatopinna	*Aviculopecten*
	Limipecten
	Streblochondria
Infaunal Bivalves	Infaunal Bivalves
Wilkingia	*Ectogrammysia*
	Edmondia
	Paleyoldia
	Phestia
	Sanguinolites
	Schizodus
	Sphenotus
Echinoderms	
crinoid and blastoid columnals	

luscs and arthropods include both stenohaline and euryhaline groups (Barnes, 1989; Dodd and Stanton, 1990; Fürsich, 1993; Hudson and others, 1995). The trilobite arthropods are thought to have been stenohaline (Robison, 1987).

The separation of stenohaline and euryhaline groups is clearly visible in the R-mode ordination plots (Figures 6 and 7) and appears to match expectations based on the previous studies cited above. Inspection of Table 1 indicates that bivalves and gastropods are very common in the Bickett Shale, whereas stenohaline groups are rare or absent in most samples, with the notable exception of B3 discussed below. The Bickett Shale samples plot separately from the samples of the two limestones in the Q-mode ordination plots (Figures 4 and 5) because of its euryhaline fauna.

Using the assumption that the Bickett Shale represents marginal marine conditions typical of bays or estuaries, the ranges of taxa in Table 1 can be used to divide the 30 taxa into stenohaline and euryhaline groups (Table 7). Note that only one Bickett sample, B3, contains bryozoans. This indicates B3 probably recorded normal marine salinity for part of its deposition, but the other typically euryhaline taxa found in B3 suggests that this sample represents time averaging of stenohaline and euryhaline taxa as salinities fluctuated from normal to brackish. Such fluctuations in salinity are typical of bays or estuaries such as San Francisco Bay (Stanton and Dodd, 1976; Dodd and Stanton, 1990, p. 358). So although sample B3 plots with the other Bickett Shale samples, it is closer to the Wymps Gap and Reynolds samples than any other Bickett samples (Figure 4).

Among all the articulate brachiopods only the genus *Orthotetes* is euryhaline as it is common in the Bickett Shale, being found in five of the seven samples. There are living euryhaline articulate brachiopods as well (Rowell and Grant, 1987). The inarticulate brachiopod *Lingula* is a well known euryhaline organism (Dodd and Stanton, 1990), but specimens were rare in the Bickett Shale and are not listed in Table 1. Among the gastropods three of the six genera, *Bullimorpha*, *Knightites*, and *Naticopsis* are stenohaline, being mostly absent from the Bickett Shale. *Naticopsis* is presumed to be

stenohaline as it is found in the Greenbrier and Reynolds limestones, and only in sample B3 from the Bickett. The epifaunal bivalve *Sulcatopinna* is also assumed to be stenohaline for the same reasons as *Naticopsis*. Nearly all infaunal bivalves were euryhaline, except for *Wilkingia*, which is absent from the Bickett Shale.

Fürsich (1993) noted that in both the Mesozoic and Cenozoic euryhaline faunas were dominated by bivalves, followed by gastropods and lingulids. Miller (1988) documented abundant and diverse bivalve assemblages in marginal marine environments beginning in the Upper Ordovician. Undoubtedly the overall taxonomic composition of euryhaline faunas has been conservative through most of the Phanerozoic based on the similarity of euryhaline organisms reported from the Ordovician (Miller, 1988) through the Mesozoic (Fürsich.1993) and into the Cenozoic (Stanton and Dodd, 1976).

Clearly the Bickett Shale contains a euryhaline fauna, which is thought to have lived in brackish rather than hypersaline conditions. There is clear evidence of fresh water influence based on the occurrence of lungfish (Elliott and Taber, 1981) and the amphibians *Mauchunkia bassa* (Hotton, 1970) and *Proterogyrinus scheelei* (Romer, 1970), organisms that lived in fresh water (Cunningham and Dickson, 1996). There is also evidence of fluvial paleo-channels and paleosols, as previously mentioned. Temporary marine conditions are indicated for sample B3 by the presence of bryozoans, plus *Naticopsis* and *Sulcatopinna*. Overall, the Bickett Shale represents deposition during a regression between the open circulation, normal marine conditions of the underlying Wymps Gap Limestone and overlying Reynolds Limestone. Presumably the clastic sediments of the Bickett were deposited in bays or estuaries along a muddy coastline where circulation with open marine conditions was restricted.

Comparison of the results of factor analysis from abundance data versus binary/taxon-count data indicates two separate groupings of stenohaline guilds within the Wymps Gap Limestone. These two groups are distinct in both the percent abundance (Table 4) and raw abundance data (Table 5) and are visible with Q-mode ordination (Figure 6), whereas for the binary/taxon-count data all the stenohaline guilds are grouped together (Table 6 and Figure 7). Factor analysis of the abundance data indicates that tethered brachiopods and bryozoans are highly correlated in their occurrence and are independent of productid brachiopods, corals, and trilobites, which are highly correlated with each other in their occurrence. Some other environmental variable(s) such as water depth or substrate type may have been responsible for these groupings. However, the separation of these two groups is subtle as it is not visible in the binary data, which has less resolving power than abundance data when recognizing paleoecologic patterns. Understanding the environmental differences between these two stenohaline groupings is beyond the scope of the present study, but we point out these two groupings as a potential subject for investigation by future workers.

In summary, evidence from the patterns of faunal distributions recorded in the bulk samples (Table 1), plus that from the factor analysis and ordination of these data (Tables 4-6, Figures 4-7), support the division of taxa into stenohaline and euryhaline groups (Table 7). The faunal divisions also support the corresponding interpretation that the Wymps Gap and Reynolds limestones record normal marine salinities, whereas the Bickett Shale records brackish water salinities. The stenohaline taxa are nearly all restricted to the Wymps Gap and Reynolds limestones where they are most diverse (Table 8). The euryhaline taxa also occur in the Wymps Gap and Reynolds limestones, but they are most diverse in the Bickett Shale (Table 8). Other evidence for paleosalinity differences between the Wymps Gap and Reynolds limestones and the Bickett Shale is sedimentological. The sedimentary facies preserved in the Wymps Gap and Reynolds limestones indicate open marine conditions suitable for stenohaline taxa (Brezinski, 1989a, b, c; Carney and Smosna, 1989). Sedimentary facies of the Bickett Shale, as previously described, record: 1) nonmarine terrestrial environments as indicated by paleosols and fluvial paleochannels; 2) freshwater aquatic environments as

indicated by the occurrence of lungfish (Elliott and Taber, 1981) and amphibians (Hotton, 1970; Romer, 1970); and 3) brackish water bays or estuaries based on the occurrence of euryhaline marine invertebrates collected during this study.

Table 8. Comparison of taxonomic diversity of stenohaline versus euryhaline taxa (Table 7) based on stratigraphic distribution (Table 1).

Stratigraphic unit and number of bulk samples	Stenohaline taxa, mean diversity	Euryhaline taxa, mean diversity
Wymps Gap Limestone (13)	9.3	4.5
Bickett Shale (7)	0.6	10.0
Reynolds Limestone (3)	7.6	2.7

CONCLUSIONS

The effect of salinity differences during deposition of the Chesterian age Wymps Gap Limestone, Bickett Shale, and Reynolds Limestone in the central Appalachian basin left a clear record in the distribution of benthic invertebrates among these units. The faunal pattern is so strong that it is visible with both abundance and binary (presence/absence) data. The Wymps Gap and Reynolds limestones record open circulation, normal marine conditions, whereas the Bickett Shale records variable salinities from fresh water to open marine associated with restricted circulation in bays or estuaries along a muddy coastline. Most articulate brachiopods were stenohaline, with the clear exception of *Orthotetes*. Bryozoans, corals, trilobites, and echinoderms were also stenohaline. The gastropods were divided between the stenohaline *Bullimorpha*, *Knightites*, and *Naticopsis* and the euryhaline *Bellerophon*, *Donaldina*, and *Straparollus*. Nearly all bivalves were euryhaline, with the exception of *Sulcatopinna* and *Wilkingia*.

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REFERENCES

Allaby, M., 1985, The Oxford Dictionary of Natural History: Oxford University Press, 688 p.

Arkle, T., Bissel, D. R., Larese, R.E., Nuhfer, E. B., Patchen, D. G., Smosna, R. A., Gillespie, W. H., Lund, R., Norton, C. W., and Pfefferkorn, H. W., 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States - West Virginia and Maryland: U. S. Geological Survey Professional Paper 1110, p. D1-D35.

Bambach, R. K., 1983, Ecospace utilization and guilds in marine communities through the Phanerozoic, *in* Tevesz, M. J. S. and McCall, P. L., eds, Biotic Interactions in Recent and Fossil Benthic Communities: Plenum Press, New York, p. 719-746.

Barnes, R. S. K., 1989, What, if anything, is a brackish-water fauna?: Transactions of the Royal Society of Edinburgh: Earth Sciences, v. 80, p. 235-240.

Berg, T. M., McInerney, M. K., Way, J. H., and MacLachlan, D. B., 1983, Stratigraphic correlation chart of Pennsylvania: Pennsylvania Geological Survey, General Geology Report 75, 1 sheet.

Brezinski, D. K., 1988, Appalachian Carboniferous trilobites: Journal of Paleontology, v. 62, p. 934-945.

Brezinski, D. K., 1989a, Late Mississippian depositional patterns in the north-central Appalachian Basin, and their implications to Chesterian hierarchal stratigraphy: Southeastern Geology, v. 30, p. 1-23.

Brezinski, D. K., 1989b, The Mississippian System in Maryland: Maryland Geological Survey, Report of Investigations No. 52, 75 p.

Brezinski, D.K., 1989c, Facies relationships and correlations between the Greenbrier of West Virginia and the Mauch Chunk of Pennsylvania, *in* Harper, J. A., ed., Geology in the Laurel Highlands of southwestern Pennsylvania: Guidebook, 54th Annual Field Conference of Pennsylvania Geologists, p.63-68.

Brezinski, D. K., 1999a, The rise and fall of Late Paleozoic trilobites of the United States: Journal of Paleontology, v. 73, p. 164-175.

Brezinski, D. K., 1999b, Chapter 9, Mississippian, *in*, Shultz, C. H., ed., The Geology of Pennsylvania: Pennsylvania Geological Survey and Pittsburgh Geological Society, p. 139-147.

Busanus, J. W., and Hoare, R. D., 1991, Bivalves (Mollusca) from the Mauch Chunk Group (Mississippian) of northern West Virginia and southwestern Pennsylvania: Journal of Paleontology, v. 65, p. 465-480.

Butts, C., 1940-1941, Geology of the Appalachian Valley in Virginia: Virginia Geological Survey Bulletin 52, Pt. 1, text, 568 p. (1940), Pt. 2, plates (1941).

Carney, C., and Smosna, R., 1989, Carbonate deposition in a shallow marine gulf, the Mississippian Greenbrier Limestone of the central Appalachian basin: Southeastern Geology, v. 30, p. 25-48.

Carter, J. L., and Carter, R. C., 1970, Bibliography and Index of North American Carboniferous Brachiopods (1898-1968): Geological Society of America, Memoir 128, 382 p.

Christopher, C. C., 1990, Late Mississippian *Girvanella*-Bryozoan mud mounds in southern West Virginia: Palaios, v. 5, p. 460-471.

Croghan, P.C., 1983, Osmotic regulation and the evolution of brackish- and fresh-water faunas: Journal of the Geological Society of London, v. 140, p. 39-46.

Cunningham, C. R., and Dickson, E. D., III, 1996, Distributions of Kansas Permo-Carboniferous vertebrate assemblages as a function of wet and dry seasons: Transactions of the Kansas Academy of Science, v. 99, nos. 1-2, p.16-28.

Davis, J. C., 1986, Statistics and Data Analysis in Geology: John Wiley and Sons, New York, 646 p.

Dennison, J. M., and Hay, W. H., 1967, Estimating the needed sampling area for subaquatic ecologic studies: Journal of Paleontology, v. 41, p. 706-708.

Dodd, J. R., and Stanton, R. J., Jr., 1990, Paleoecology, Concepts and Applications: John Wiley and Sons, New York, 502 p.

Droser, M. L., Bottjer, D. J., and Sheehan, P. M., 1997, Evaluating the ecological architecture of major events in the Phanerozoic history of marine invertebrate life: Geology, v. 25, p. 167-170.

Droser, M. L., Bottjer, D. J., Sheehan, P. M., and McGhee, G. R., Jr., 2000, Decoupling of taxonomic and ecologic severity of Phanerozoic marine mass extinctions: Geology, v. 28, p. 675-678.

Elliott, D. K., and Taber, A. C., 1981, Mississippian vertebrates from Greer, West Virginia: Proceedings of the West Virginia Academy of Science, v. 53, p. 14.

Etter, W., 1999, Community analysis, *in*, Harper, D. A. T., ed., Numerical Palaeobiology: John Wiley and Sons, Chichester, p. 285-360.

Flint, N. K., 1965, Geology and mineral resources of southern Somerset County, Pennsylvania: Pennsylvania Geological Survey, County Report 56A, 267 p.

Fürsich, F. T., 1993, Palaeoecology and evolution of Mesozoic salinity-controlled benthic macroinvertebrate associations: Lethaia, v. 26, p. 327-346.

Gauch, H. G., Jr., 1982, Multivariate Analysis in Community Ecology: Cambridge University Press, 298 p.

Girty, G. H., 1923, Observations on the faunas of the Greenbrier Limestone and adjacent rocks, *in*, Reger, D.B., Tucker County, West Virginia Geological Survey Report, p. 450-488.

Henry, T. W., and Gordon, M., Jr., 1992, Middle and upper Chesterian brachiopod biostratigraphy, eastern Appalachians, Virginia and West Virginia: Oklahoma Geological Survey Circular 94, 21 p.

Hickok, W. O., and Moyer, F. T., 1940, Geology and mineral resources of Fayette County, Pennsylvania: Pennsylvania Geological Survey, Fourth Series, Bulletin C26, 530 p.

Hoare, R. D., 1993, Mississippian (Chesterian) bivalves from the Pennsylvanian stratotype area in West Virginia and Virginia: Journal of Paleontology, v. 67, p. 374-396.

Horowitz, A. S., and Rexroad, C. B., 1972, Conodont biostratigraphy of some United States Mississippian sites: Journal of Paleontology, v. 46, p. 884-891.

Hoskins, D. M., Inners, J. D., and Harper, J. A., 1983, Fossil Collecting in Pennsylvania: Pennsylvania Geological Survey, General Geology Report 40, 215 p.

Hotton, N., III, 1970, *Mauchunkia bassa*, gen. et sp. nov., an anthracosaur (Amphibia, Labyrinthodonta) from the Upper Mississippian: Kirtlandia, v. 12, p. 1-38.

Hudson, J.D., Clements, R. G., Riding, J. B., Wakefield, M. I., and Walton, W., 1995, Jurassic paleosalinities and brackish-water communities - a case study: Palaios, v. 10, p. 392-407.

Kelleher, G. T., and Smosna, R., 1993, Oolitic tidal-bar reservoirs in the Mississippian Greenbrier Group of West Virginia, *in* Keith, B. D., and Zuppann, C. W., eds., Mississippian Oolites and Modern Analogs: American Association of Petroleum Geologists, Studies in Geology 35, p. 163-173.

Lake, A. M., 1998, Marine paleoecology of Upper Mississippian strata in northern West Virginia and southwestern Pennsylvania: West Virginia University, Master's thesis, 95 p.

Lucke, J. B., 1939, Limestones from Mississippian to Permian, *in* McCue, J. B., Lucke, J. B., and Woodward, H. P., eds, Limestones of West Virginia: West Virginia Geological Survey, v. 12, p. 27-72.

Maples, C. G., and Waters, J. A., 1987, Redefinition of the Meramecian/Chesterian boundary (Mississippian): Geology, v. 15, p. 647-651.

Meek, F. B., 1871, Lists of Carboniferous fossils from West Virginia, descriptions of new species: West Virginia University Board of Regents Report, v. 3, p. 68-73.

Miller, A. I., 1988, Spatio-temporal transitions in Paleozoic Bivalvia: an analysis of North American fossil assemblages: Historical Biology, v. 1, p. 251-273.

Oliver, W. A., Jr., and Coates, A. G., 1987, Phylum Cnidaria, *in* Boardman, R. S., Cheetham, A. H., and Rowell, A. J., eds., Fossil Invertebrates: Blackwell Scientific Publications, Palo Alto, p. 140-193.

Pearse, A. S., and Gunter, G., 1957, Salinity, *in* Hedgpeth, J. W., ed., Treatise on Marine Ecology and Paleoecology, Volume 1, Ecology: Geological Society of America, Memoir 67, p. 129-158.

Peel, J. S., 1987, Class Gastropoda, *in* Boardman, R. S., Cheetham, A. H., and Rowell, A. J., eds., Fossil Invertebrates: Blackwell Scientific Publications, Palo Alto, p. 304-329.

Pojeta, J., Jr., 1987, Class Pelecypoda, *in* Boardman, R. S., Cheetham, A. H., and Rowell, A. J., eds., Fossil Invertebrates: Blackwell Scientific Publications, Palo Alto, p. 386-435.

Reger, D. B., 1926, Mercer, Monroe, and Summers Counties: West Virginia Geological Survey, 963 p.

Rexroad, C. B., and Clarke, C. E., 1960, Conodonts from the Glen Dean Formation of Kentucky and equivalent for-

mations of Virginia and West Virginia: Journal of Paleontology, v. 34, p. 1202-1206.

Robison, R. A., 1987, Superclass Trilobitomorpha, *in* Boardman, R. S., Cheetham, A. H., and Rowell, A. J., eds., Fossil Invertebrates: Blackwell Scientific Publications, Palo Alto, p. 221-241.

Rollins, H. B., and Brezinski, D. K., 1988, Reinterpretation of crinoid-platycerid interaction: Lethaia, v. 21, p. 207-217.

Romer, A. S., 1969, A temnospondylus labyrinthodont from the Lower Carboniferous: Kirtlandia, v. 10, p. 1-16.

Romer, A. S., 1970, A new anthracosaurian labyrinthodont, *Proterogyrinus scheelei*, from the Lower Carboniferous: Kirtlandia, v. 10, p. 1-16.

Rowell, A. J., and Grant, R. E., 1987, Phylum Brachiopoda, *in* Boardman, R. S., Cheetham, A. H., and Rowell, A. J., eds., Fossil Invertebrates: Blackwell Scientific Publications, Palo Alto, p. 445-496.

Simonsen, A. H., 1981, Paleoecology of fenestrate bryozoans in the Wymps Gap Limestone of southwestern Pennsylvania: Pennsylvania Geology, v. 12, no. 2, p. 11-15.

Simonsen, A. H., 1988, Chesterian (Mississippian) fossil site in the Mauch Chunk Formation, Chalk Hill, Pennsylvania: Northeastern Geology v. 10, p. 50-56.

Smith, J., 1999, Multivariate techniques in palynofacies analysis, *in* Harper, D. A. T., ed., Numerical Palaeobiology: John Wiley and Sons, Chichester, p. 361-393.

Smosna, R., and Koehler, B., 1993, Tidal origin of Mississippian oolite on the West Virginia Dome, *in* Keith, B. D., and Zuppann, C. W., eds., Mississippian Oolites and Modern Analogs: American Association of Petroleum Geologists, Studies in Geology, v. 35, p. 149-162.

Stanton, R. J., Jr., and Dodd, J. R., 1976, The application of trophic structure of fossil communities in paleoenvironmental reconstruction: Lethaia, v. 9, p. 327-342.

Swan, A. R. H., and Sandilands, M., 1995, Introduction to Geological Data Analysis: Blackwell Science, Oxford, 446 p.

Thein, M. L., and Nitecki, M. H., 1974, Chesterian (upper Mississippian) Gastropoda of the Illinois Basin: Fieldiana Geology, v. 34, p. 1-238.

Thomson, K. S., 1966, On the relationships of certain Carboniferous Dipnoi, with descriptions of four new forms: Royal Society of Edinburgh, Proceedings (1964-1966), sec. B., v. 69, p. 221-245.

Tilton, J. L., 1928, Geology from Morgantown to Cascade, West Virginia, along state route number 7: Proceedings of the West Virginia Academy of Science, v. 2, p. 1-22.

Yielding, C. A., and Dennison, J. M., 1986, Sedimentary response to Mississippian tectonic activity at the east end of the 38th Parallel fracture zone: Geology, v. 14, p. 621-624.

APPENDIX 1. LOCALITY INFORMATION (FIGURES 2 AND 3).

Greer Limestone Company -- Active limestone quarry/mine along Deckers Creek, Masontown Quadrangle, Monongalia County, West Virginia. UTM coordinates: Zone 17; 599,200 m E; 4,380,875 m N (Busanus and Hoare, 1991, locality 2).

Buckeye Stone Company -- Active limestone quarry in Blaney Hollow, Lake Lynn Quadrangle, Monongalia County, West Virginia. UTM coordinates: Zone 17; 604,000 m E; 4,394,900 m N (Busanus and Hoare, 1991, locality 4).

Ryan Hollow-- Abandoned limestone quarry near the top of Ryan Hollow, Lake Lynn Quadrangle, Monongalia County, West Virginia. UTM coordinates: Zone 17; 605,580 m E; 4,396,000 m N.

Lake Lynn Laboratory -- Former limestone quarry/mine, operated by National Institute of Occupational Safety and Health (NIOSH), in Dragoo Hollow, Lake Lynn Quadrangle, Fayette County, Pennsylvania. UTM coordinates: Zone 17; 605,570 m E; 4,397,500 m N (Busanus and Hoare, 1991, locality 5).

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SEDIMENTOLOGY, PALEOECOLOGY, AND SEQUENCE STRATIGRAPHY OF THE WINIFREDE SHALE, KANAWHA FORMATION (MID-PENNSYLVANIAN), SOUTHERN WEST VIRGINIA

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ABSTRACT

The Winifrede Shale and genetically related, overlying sandstone lithofacies are 4.3-7.3 m thick and contain five sedimentary facies: (1) sandy mudstone and shale containing brachiopods, bivalves, gastropods, crinoids, bryozoans, and nautiloids; (2) shale distinguished by highly comminuted shells aligned in mm-thick laminae and carbonaceous shale at the top; (3) brachiopod-dominated shale which becomes sideritic in the upper part; (4) fossiliferous siderite bed; and (5) faunally barren, coarsening-upward, shale and sandstone. These facies accumulated during part of a fourth-order sea level cycle.

Facies 1 and 2 represent a condensed, transgressive systems tract bounded at its base by a ravinement surface and at its top by a maximum flooding surface. The carbonaceous shale developed under dysaerobic conditions and contains *Zoophycos* and laminae deposited as distal tempestites. Facies 3-5 represent a regressive facies sequence and comprise the highstand systems tract. Centimeter-thick shell hashes in facies 3 accumulated as proximal tempestites, and alternate with well-preserved shell beds buried by rapid mud deposition. The appearance and increasing frequency of siderite nodules and beds reflect fluctuating or decreased salinity associated with more nearshore conditions. Continued regression subsequently resulted in deposition of a faunally barren, coarsening upward, delta front sequence (facies 5). Taphonomic attributes and inferred ecologic conditions of body and trace fossils are instrumental in recognition of transgressive and highstand systems tracts.

The levels of boron, chromium, and nickel in the clay fraction of the shales are, on average, higher in facies 1-3 where stenohaline taxa are prevalent, and generally lower where shales become sideritic and fossil invertebrates are sparse or absent. The results indicate that geochemical analyses using these elements may prove useful in paleosalinity interpretations for other Middle Pennsylvanian shales of suspected marine origin in the central Appalachian Basin.

INTRODUCTION

This paper evaluates the lithology, sedimentology, and paleontology of the Winifrede Shale Member based on detailed analyses of three outcrops its type area. These outcrops occur along an 11 km, east-west transect in Kanawha County, West Virginia (Figure 1).

The Winifrede Shale is one of fourteen marine shales that are currently recognized in the Kanawha Formation of southern West Virginia (Blake, 1998; Figure 2). It is equivalent to the Magoffin Shale of eastern Kentucky (Wanless, 1939; Outerbridge, 1976; Blake and others, 1994), which together comprise one of the most distinctive and widespread marine deposits in the central Appalachian Basin. Bennington (1996, 1998) described and analyzed the stratigraphy and biofacies in the Magoffin Shale. However, little detailed work has been

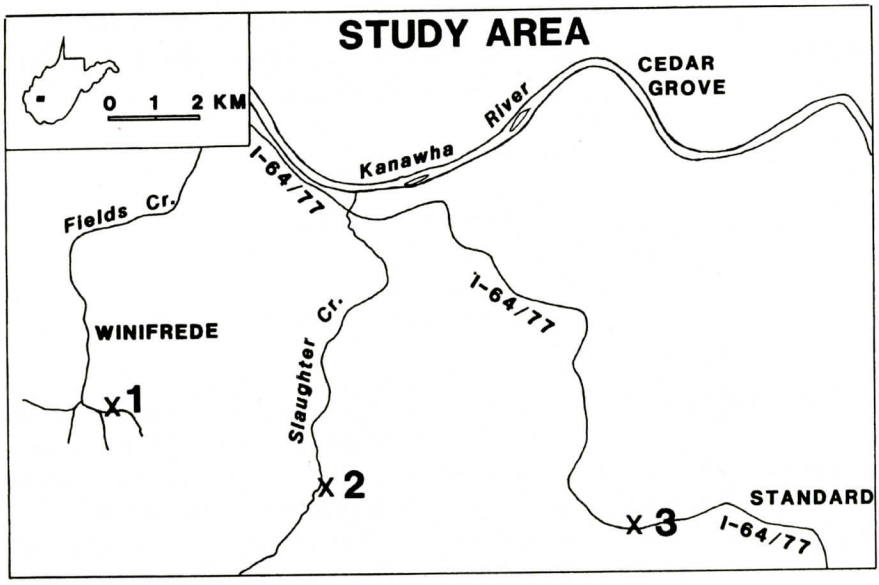

Figure 1. Location map showing outcrop localities 1, 2, and 3 in southern Kanawha County, West Virginia (Belle and Cedar Grove 7.5' quadrangles).

done on the invertebrate paleontology in West Virginia. Previous paleontological studies have been limited to listings of macrofaunal taxa and generalized paleoenvironmental interpretations (Price, 1914; Henry and Gordon, 1979). Henry and Gordon (1979) suggested an Atokan age for the Winifrede Shale based on marine invertebrates in its type area. Megafloral correlations indicate a latest Westphalian B age (Gillespie and Pfefferkorn, 1979). The Winifrede Shale has been correlated with the Aegir marine band which marks the boundary between the Duckmantian and Bolsovian Series in Europe (Riley and Turner, 1994; Wagner and Winkler Prins, 1991).

Subtle internal variations within shallow marine, offshore mudstone sequences may reflect periodic variations in dissolved oxygen, near-bed energy level, turbidity, and substrate conditions (Johnson and Baldwin, 1986). In addition, recognition of depth-related changes in sedimentologic and paleontologic parameters is

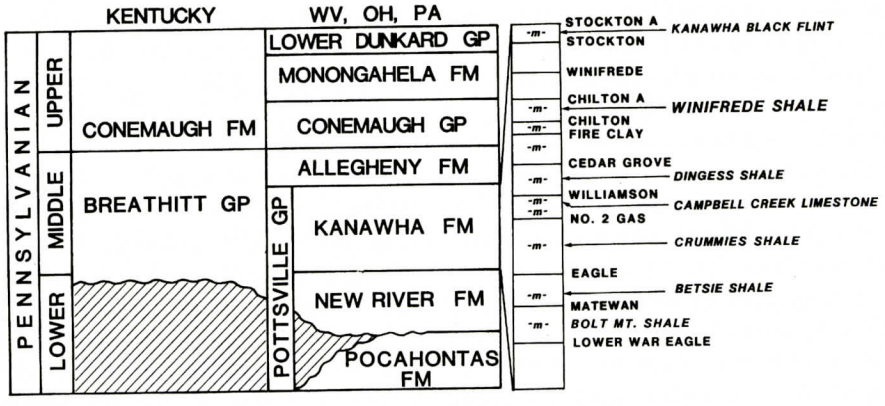

Figure 2. Stratigraphic framework for the Winifrede Shale. Selected marine shales (italics) and coal beds from the Kanawha Formation are shown in the expanded section at the right (modified from Blake, 1998, and Martino, 1996).

crucial in deciphering the anatomy of sequences and parasequences (Brett, 1998).

The objectives of this paper are to (1) delineate and interpret sedimentary facies and associated faunal assemblages within the Winifrede Shale and overlying strata; (2) recognize systems tracts and bounding stratal surfaces, and (3) provide a geochemical profile through these facies and determine if trace elements including boron, chromium, copper, gallium, nickel and vanadium in the clay fraction show trends that may reflect paleosalinity.

Methods

Invertebrates were found to occur over a 1.5-3 m stratigraphic interval at the base of the Winifrede Shale. This interval was incrementally sampled at each outcrop by removing a continuous vertical sequence of blocks with lateral dimensions of approximately 30 cm x 30 cm. The vertical limits for each sample ranged from 7 to 30 cm and were determined by bedding planes marking lithologic changes. A maximum thickness of 30 cm was used for the sample increments through thicker facies. The sample blocks were then dissected in the laboratory where taxa were identified and their relative abundance noted.

A parallel set of incremental samples was also geochemically analyzed to evaluate the content of salinity-sensitive trace elements. The samples were crushed, carbonate removed using hydrochloric acid, and clay content segregated by settling methods. The resulting clay samples were analyzed by emission spectroscopy for boron, copper, chromium, gallium, nickel and vanadium.

RESULTS

Sedimentary Facies

Stratigraphic sections through the Winifrede Shale at localities 1, 2, and 3 are shown in Figures 3-5. Component sedimentary facies and the vertical distribution and relative abundance of fossils are also indicated. The Winifrede Shale at all three localities has a sharp, erosional contact with an underlying coal bed or underclay. This contact is overlain by five sedimentary facies which will be described here in ascending order.

Facies 1: Basal Sandy Mudstone and Shale

The Winifrede Shale contains a highly fossiliferous, basal interval 10-25 cm thick (Figures 6A,6B,7A,7B). This interval consists of dark gray, highly calcareous, sandy mudstone that grades upward into shale. The term mudstone, as used here, refers to a blocky or massive, fine grained rock which contains approximately equal amounts of clay and silt-sized grains, whereas shale refers to a laminated, indurated rock with more than two-thirds clay-sized minerals (Jackson, 1997). Thin, shell-rich zones up to 1 cm thick are interlayered with sparsely fossiliferous mudstone. This basal interval is packed with bivalves, crinoids, and brachiopods dominated by *Anthracospirifer*. Other less abundant taxa include small gastropods, fenestellid bryozoans, nautiloids, and ostracodes. Very fine quartz sand is abundant in some layers and coarser flakes of muscovite are present. Coal and mud rip-ups clasts are common within the sandy mudstone. Sorting is moderate in some sandy layers and generally decreases upward through facies 1. Quartz sand content and shell disarticulation are highest at the base and decrease upward. Convex-upward shells predominate at the base, whereas concave-upward shells become more frequent in the middle and upper parts of facies 1.

Facies 2: Dark Gray and Black Calcareous Shale

Facies 2 has a gradational contact with facies 1 and consists of a 20-37 cm thick shale. This facies is distinguished by size-sorted fossil fragments and highly carbonaceous character in the upper 7-15 cm. Well-sorted coquinas consisting mainly of 1-3 mm bioclasts of ostracodes, brachiopod spines, and shell fragments are interlaminated with highly carbonaceous shale with diffuse, comminuted skeletal material. Very rare gastropods, nuculid bivalves, black phosphate nodules also occur within the coquinas.

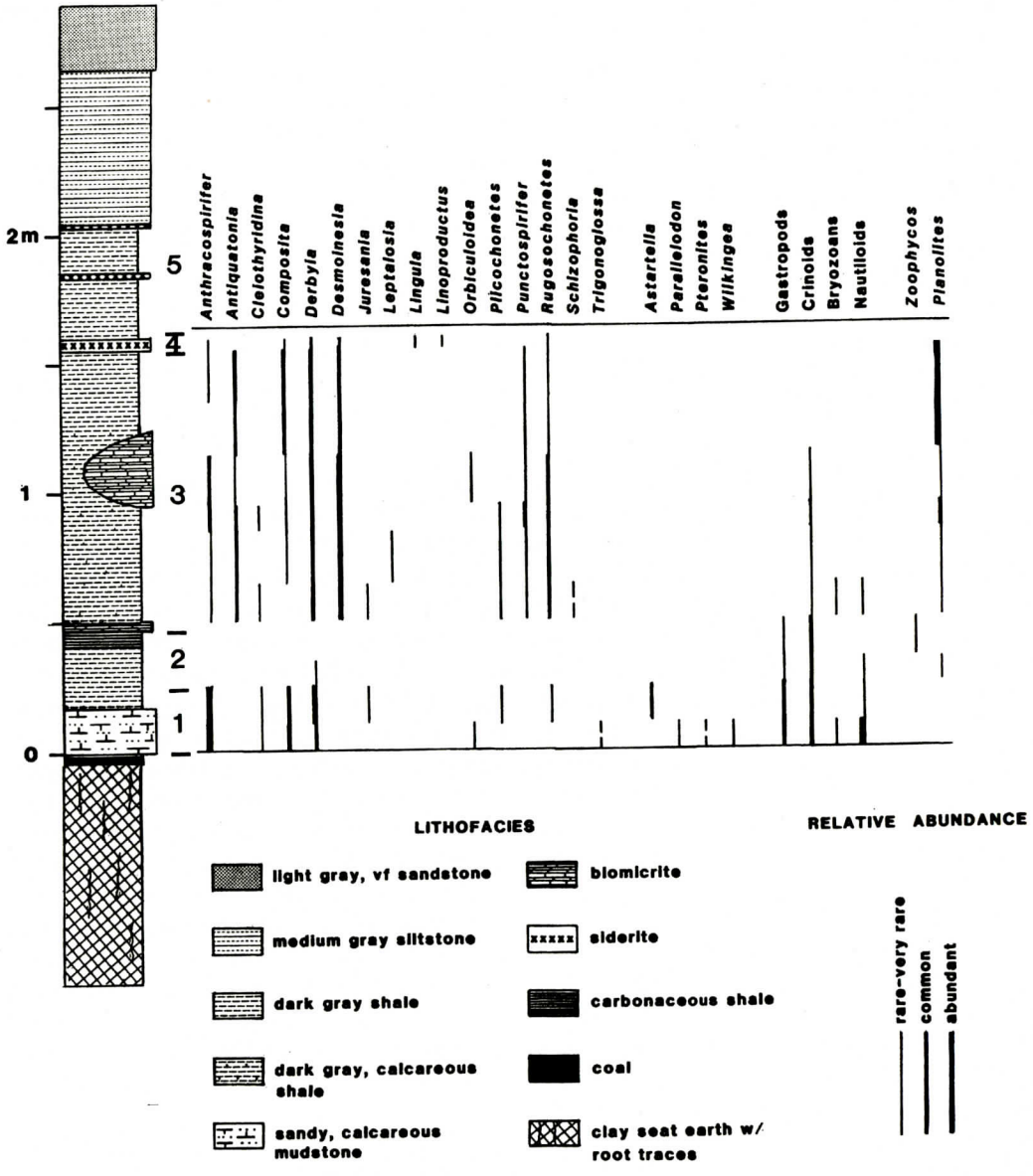

Figure 3. Stratigraphic section at locality 1 showing component sedimentary facies in the Winifrede Shale, and the vertical distribution and relative abundance of body and trace fossils. Dashed vertical lines indicate uncertainty in identity of genera due to poor preservation. Facies 1-5 are shown to the right of stratigraphic column (modified from Martino, 1996).

The trace fossil *Zoophycos* is present in the carbonaceous shale. *Planolites* also occurs locally as straight to sinuous tubes 1.5-2 mm in diameter. In some places the burrow fill is packed with pellets about 0.25 mm in diameter. Many of the *Planolites* burrows are pyritized.

Facies 3: Dark Gray Calcareous/ Sideritic Shale

Facies 3 consists of dark gray, fossiliferous shale 1.1-2.4 m thick. Biomicrite occurs locally as thin beds, laminae, and as limestone nodules up to 30 cm thick and 80 cm wide. Cone-in-cone structures are common in the nodules. Sid-

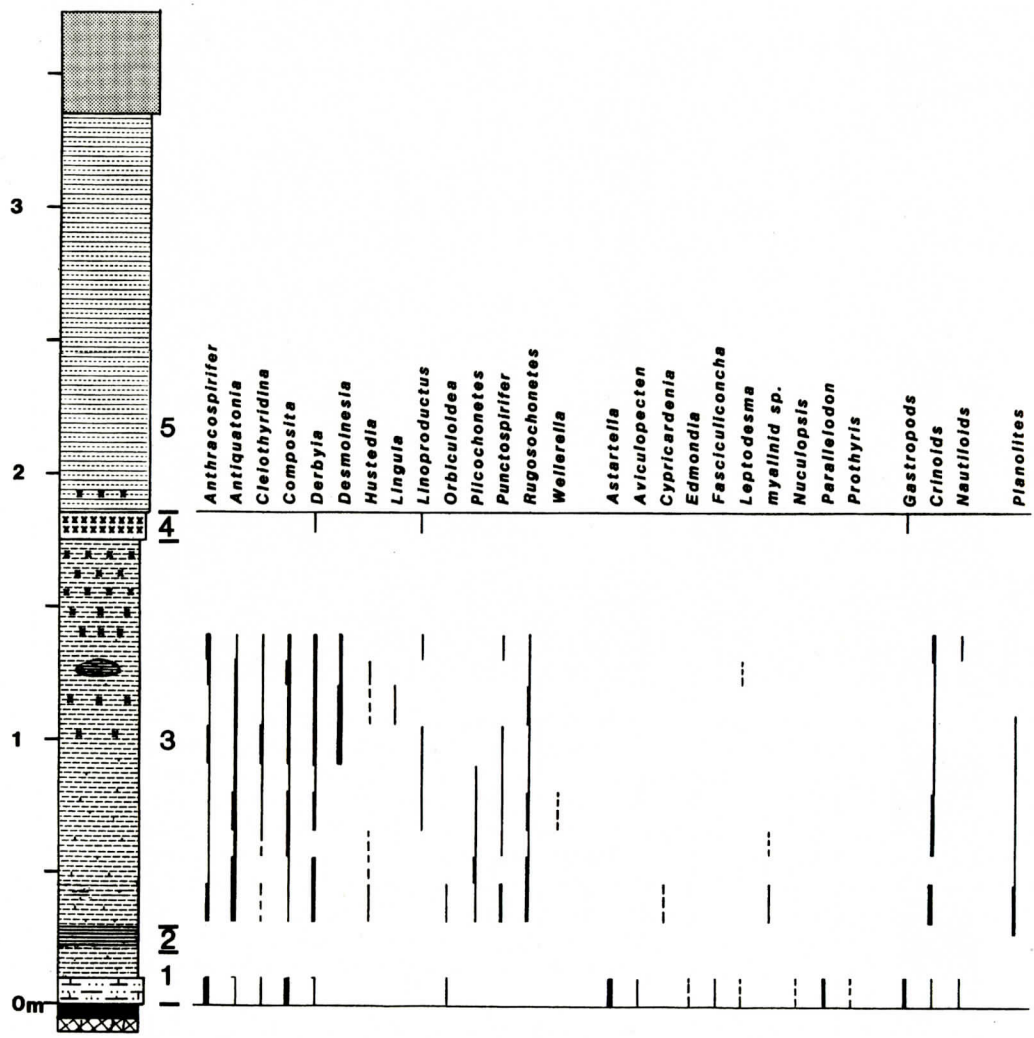

Figure 4. Stratigraphic section at locality 2 showing component sedimentary facies in the Winifrede Shale and the vertical distribution and relative abundance of body and trace fossils.

erite nodules appear in the upper part of the facies and increase in abundance toward the top. This facies contains three divisions based on the character of the fauna. These include in ascending order: (1) well-sorted coquinas that alternate with whole shell beds; (2) well-preserved, abundant shells with minimal sorting or fragmentation; (3) decreasing quality of shell preservation except in siderite nodules or where replaced by siderite.

Articulate brachiopods predominate, especially *Anthracospirifer, Composita, Antiquatonia, Desmoinesia*, and *Derbyia*. Crinoid ossicles are fairly common in the lower and middle portions of facies 3. Bivalves, nautiloids, and bryozoans are scarce and restricted to the lowermost portion of facies 3 (Figures 3-5).

A distinctive, crinoidal packstone 3 cm thick occurs at the base of facies 3 at localities 1 and 3. Allochems are granule to coarse-sand sized, show moderate sorting and size grading, and include a mixture of crinoid plates, brachiopod shells and spines, and phosphate granules (Figure 7D).

Within the lower portion of facies 3, coquinas alternate with poorly sorted shell beds

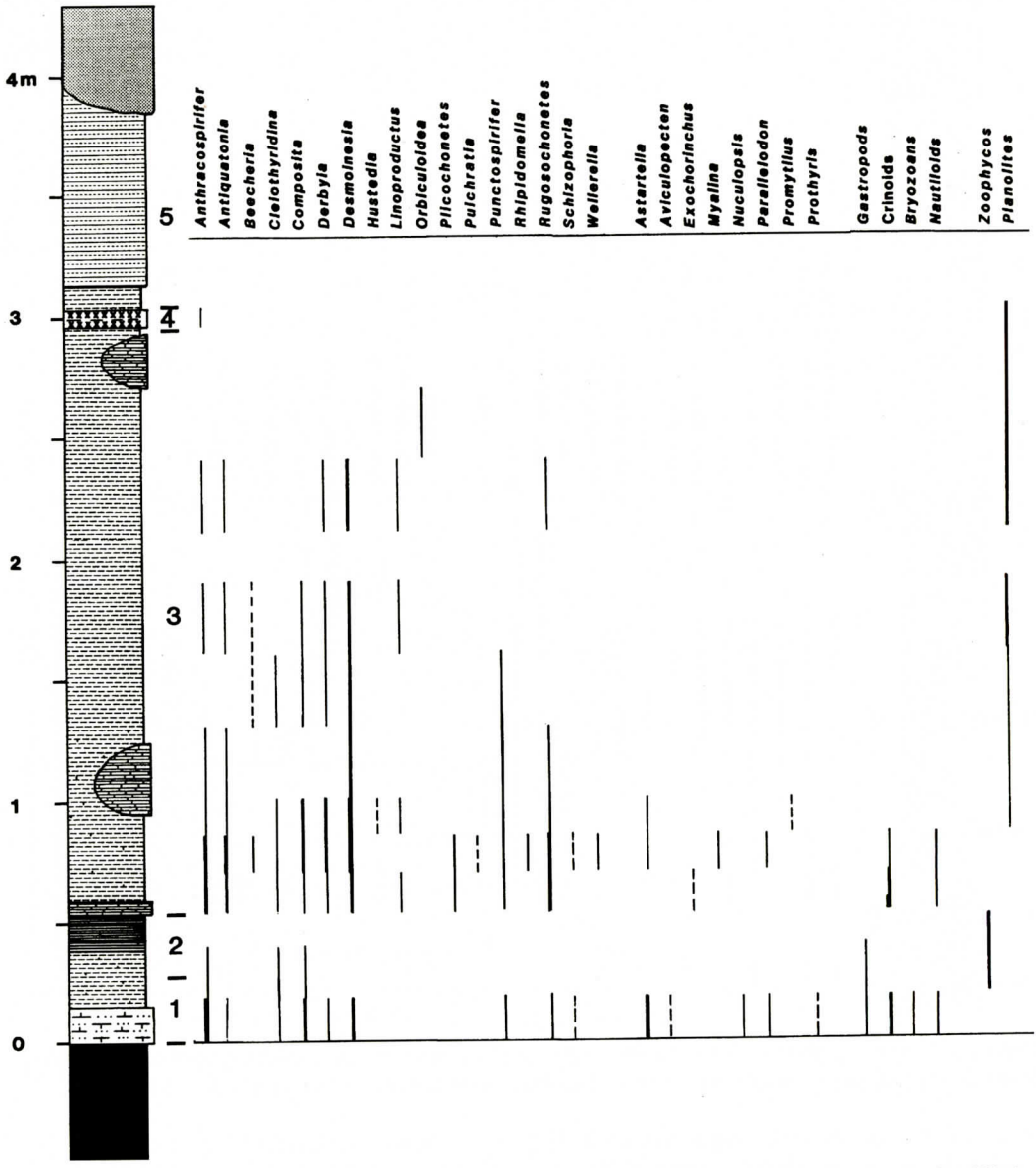

Figure 5. Stratigraphic section at locality 3 showing component sedimentary facies in the Winifrede Shale and the vertical distribution and relative abundance of body and trace fossils.

consisting mainly of articulated brachiopods (e.g. *Antiquatonia, Composita*, chonetids) and crinoid stem segments. An example of the latter occurs at locality 2 where four 1 cm thick layers of articulated brachiopods alternate with coquinas 2-3 cm thick. Faunally barren shale intervals are also intermittently developed within facies 3.

In the limestone nodules, the disruptive growth of calcite crystals in cone-in-cone structures has shattered the calcareous shale of facies 3 into a diagenetic breccia. "Clasts" of shale occur floating in calcitic host material. Brachiopods within the limestone nodules are often partly or completely replaced with pyrite.

Facies 4: Siderite Bed

A laterally persistent bed of siderite 4-10 cm

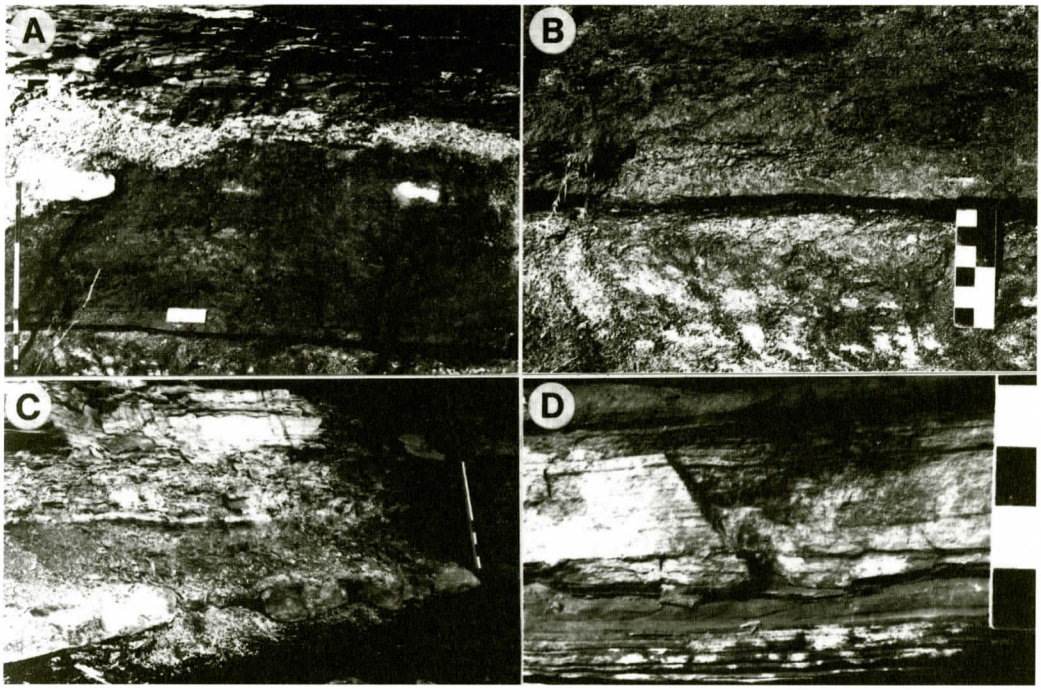

Figure 6. Winifrede Shale facies, locality 1. Jacobs staff = 1.5 m. Bar scale = 30 cm, with 15 cm and 5 cm divisions. A) basal sandy mudstone of facies 1 forms a thin ledge (beneath bar scale) above underclay of Chilton Rider coal; limestone nodules are visible in facies 3 near middle of photo (at top of Jacobs staff); lower part of parallel laminated, interbedded sandstone and shale interval (facies 5) is shown at top of photo; B) scale resting on underclay; its contact with overlying ledge-forming fossiliferous, sandy mudstone (facies 1) is interpreted as a ravinement surface; C) limestone nodules in calcareous shale near middle of facies 3 in lower part of photo; a thin siderite bed (facies 4) occurs as white band near center of photo; thinly interbedded sandstone and shale of facies 5 is shown in upper part of photo; D) parallel-laminated very fine sandstone with minor scour at base, interbedded with shale (facies 5).

thick is present at localities 1-3 (Figure 6C). The bed contains a well-preserved, autochthonous fauna including articulated calcareous and phosphatic brachiopods, rare bivalves, gastropods, and phosphatic fish bone fragments, teeth, and scales. Internally, the siderite bed is structureless.

Shale within facies 3 directly below the siderite bed contains very few fossils at localities 2 and 3 (Figures 4, 5). At locality 1, seven brachiopod genera persist through facies 3 to its base (Figure 3).

Facies 5: Coarsening-Upward, Shale-Sandstone Facies

The top of the Winifrede Shale is transitional with a faunally barren, coarsening-upward sequence that is capped by underclay and coal (Figure 6A, C, D). This stratigraphic interval is 3.5-5.6 m thick at localities 1 and 2. At locality 3, facies 5 is missing and the Winifrede Shale is truncated by a multistory, channel sandstone.

Facies 5 consists of dark gray shale, medium gray siltstone, and light gray, very fine to fine sandstone. The sandstone is thin-bedded with parallel lamination, and contains coarse mica and plant detritus (Figure 6D). Convolute lamination is locally developed near the base of facies 5.

Trace Element Geochemistry

Forty-eight incremental samples from localities 1-3 were analyzed to determine the concen-

Table 1. Trace element concentrations (ppm) in clay fraction of incremental samples taken from Winifrede Shale Member at localities 1, 2, and 3. Intervals shown on left side refer to the height of the sample interval above the base of the Winifrede Shale, defined by the top of the Chilton rider coal or its seat earth. Sid.= siderite bed or nodules.

Locality 1 Interval (cm)	B	Cr	Cu	Ga	Ni	V	Facies
0-10	100	300	20	70	50	150	1
10-25	50	100	20	30	70	100	
25-50	100	500	30	70	50	150	2
50-65	100	200	30	70	100	150	3
65-95	50	150	20	50	70	70	
95-115	70	150	30	50	70	150	
115-130	70	200	50	50	100	150	
130-152	70	200	30	50	70	150	
152-155 (sid.)	50	100	15	50	30	70	4
155-185	50	70	30	30	30	100	5
185-215	30	100	30	30	50	70	
215-245	50	200	50	50	70	100	
245-275	50	100	30	50	50	100	
275-300	50	150	100	70	50	100	
Locality 2	**B**	**Cr**	**Cu**	**Ga**	**Ni**	**V**	**Facies**
0-10	70	200	50	70	70	100	1
10-20	50	150	20	50	100	100	
20-35	70	150	30	50	100	150	2
35-55	70	150	30	50	70	100	3
55-80	50	70	20	30	30	70	
80-110	70	100	20	30	50	100	
110-130	70	150	30	50	70	150	
130-140	50	100	20	30	50	70	
140-150[sid.n.]	50	100	30	50	70	70	
150-165	30	150	20	50	50	100	
165-175	30	100	30	50	30	100	
175-185 (sid.)	50	150	50	50	50	150	4
185-200	30	100	30	50	50	150	5
200-225	50	100	30	50	50	150	
225-250	50	100	30	50	50	150	
250-275	50	100	50	50	50	150	
Locality 3	**B**	**Cr**	**Cu**	**Ga**	**Ni**	**V**	**Facies**
0-15	70	70	15	50	30	100	1
15-25	50	150	20	30	70	100	
25-40	70	100	20	50	70	150	2
40-52	70	150	20	50	70	150	
52-55	70	200	20	50	100	150	3
55-70	70	200	50	50	100	150	
70-85	50	150	20	30	50	150	
85-100	50	100	30	50	70	150	
100-130	50	100	30	50	50	70	
130-160	50	100	30	50	50	100	
160-190	70	150	30	50	70	150	
190-210	70	70	20	30	50	100	
210-240	70	100	20	30	70	150	
240-270	70	100	20	50	70	100	
270-295	70	100	15	30	70	100	
295-303 (sid.)	30	70	20	30	30	70	4
303-333	50	70	20	50	30	70	5
333-363	50	150	50	50	70	100	

Table 2. Average trace element concentrations (ppm) from the clay fraction of incremental samples in facies 1-3 compared with averages for facies 4 and 5. Ranges of values are shown in parentheses. B, Cr, and Ni have concentrations that differ between facies 1-3 and 4-5.

Locality 1	B	Cr	Cu	Ga	Ni	V
Facies 1-3	76.3	225	28.8	55	72.5	133.8
0-152 cm	(50-100)	(100-300)	(20-50)	(30-70)	(50-100)	(70-150)
Facies 4-5	46.7	120	42.5	46.7	46.7	90
152-300 cm	(30-50)	(70-120)	(15-100)	(30-70)	(30-70)	(70-100)
Locality 2	**B**	**Cr**	**Cu**	**Ga**	**Ni**	**V**
Facies 1-3	55.4	129	27.3	46.4	62.7	100.9
0-175 cm	(30-70)	(70-200)	(20-50)	(30-70)	(30-100)	(70-150)
Facies 4-5	46	110	38	50	50	150
175-275 cm	(30-50)	(100-150)	(30-50)	(50)	(50)	(150)
Locality 3	**B**	**Cr**	**Cu**	**Ga**	**Ni**	**V**
Facies 1-3	63.3	122.6	24	43.3	66	124.7
0-295 cm	(50-70)	(70-200)	(15-50)	(30-50)	(30-100)	(70-150)
Facies 4-5	43.3	96.6	30	43.3	43.3	80
295-363 cm	(30-50)	(70-150)	(20-50)	(30-50)	(30-70)	(70-100)

tration and vertical variability of six trace elements that may be potential paleosalinity indicators. The data from these analyses are shown in Tables 1 and 2.

DISCUSSION

Interpretation of Sedimentary Facies

Facies 1

The sharp contact between the basal bed and underlying coal and underclay is interpreted as a ravinement surface. It represents an erosional disconformity formed as coastal marine currents reworked sediments during the early stages of transgression (Swift, 1968). A ravinement surface occurs at the base of most of the marine shales in the Kanawha Formation (Martino, 1994).

The predominance of epifaunal, suspension-feeding, stenohaline taxa (articulate brachiopods, crinoids) and the high diversity of the assemblage indicates an open marine, clear-water setting. It is likely that much of the terrigenous sediments were trapped in estuaries during the initial stages of the Winifrede transgression and prevented from reaching shallow shelf areas (Martino, 1994, 1996).

Although the taxa show considerable disarticulation and moderate sorting at the base of facies 1, evidence for transport (e.g. shell fragmentation, size-sorting, disarticulation, concave-upward orientation) decreases upward (Figure 7B). The character of the fauna in the upper portions of the ravinement bed indicates that they are autochthonous. These attributes, along with an upward decrease in sand, reflect increasing distance from shore. This was accompanied by less frequent and/or less intense turbulence associated with greater water depth. In the middle and upper parts of facies 1, episodes of rapid mud deposition occasionally occurred. This caused rapid burial and minimal shell fragmentation and distarticulation that is evident in some horizons. In contrast, prolonged exposure would have favored development of beds with an abundance of fragmented and disarticulated skeletal elements (Speyer and Brett, 1988). A similar bed with autochthonous brachiopods and bivalves overlying a ravinement surface occurs at the base of the Lower Pennsylvanian Morris Shale of Alabama (Lui and Gastaldo, 1989; Gastaldo and others, 1991).

The *Anthracospirifer*-dominated assemblage of facies 1 has similarities with the moderate-water-depth, stenotopic faunal association of Boardman and others (1984), the offshore marine faunal association of Ferm and Williams (1965), and the nearshore brachiopod biofacies of Bennington (1996). The term faunal association refers to a consistently recurring group of

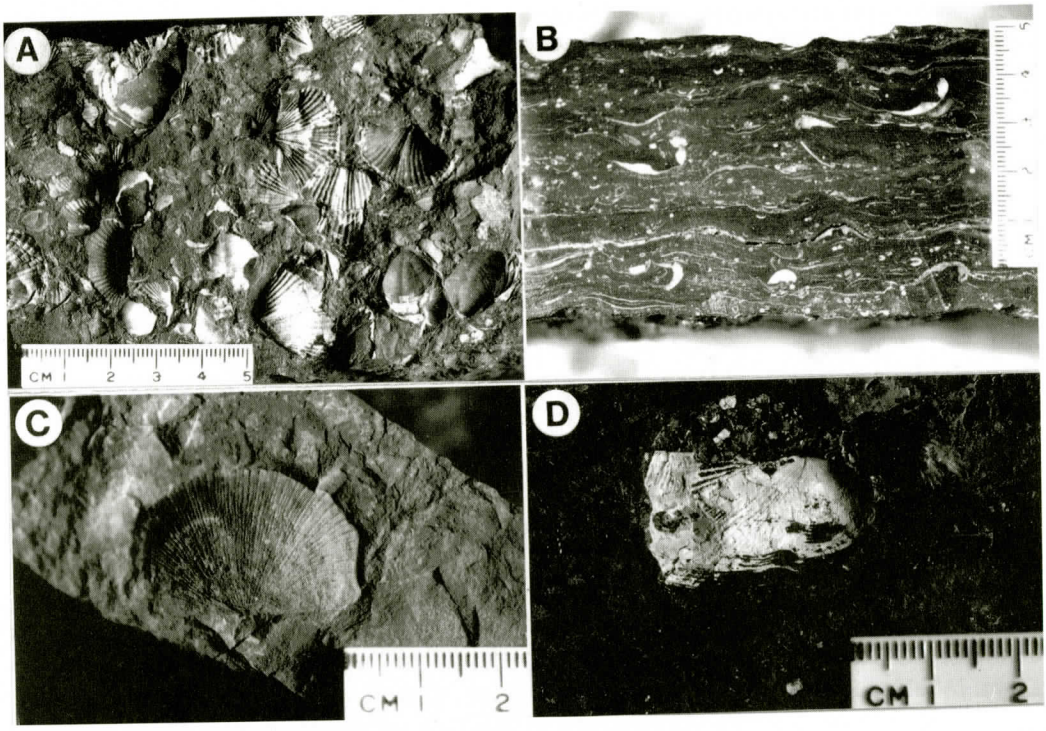

Figure 7. A) basal sandy mudstone (facies 1), view of upper surface of parting within bed, with abundant thick-shelled brachiopods, mainly *Anthracospirifer* with a few *Composita*; B) vertical cross-section through basal sandy mudstone showing brachiopods and crinoid ossicles; darkening upward is due to decreasing quartz sand content; C) *Derbyia*, from siderite bed (facies 4); D) *Parallelodon*, from basal sandy mudstone (facies 1).

fossils of organisms that presumably lived together, but which represent only part of a paleocommunity (Goldring, 1991). Biofacies, as used here, refers to a distinctive assemblage of organisms formed under a particular set of environmental parameters; it is equivalent to an ecologic association of fossils (Jackson, 1997).

Anthracospirifer dominates in nearshore facies of the Pennsylvanian Minturn Formation of Colorado, comprising 50% or more of the brachiopod fauna from the shoreline to an interpreted water depth of 6 m (Stevens, 1971). However, the high diversity of taxa (i.e. taxonomic richness, Dodd and Stanton, 1990) in facies 1 (including brachiopods, bivalves, gastropods, cephalopods, bryozoans, and crinoids) is not likely to be developed in a eurytopic nearshore setting. It is more likely that the fauna of the basal bed represents time-averaged sample (Fursich, 1978; Dodd and Stanton, 1990) of a nearshore and an offshore faunal association. The nearshore, clear-water faunal association was dominated by thick-shelled brachiopods like *Composita* and *Anthracospirifer* and developed during the initial stages of the transgression. The offshore association contains brachiopods better adapted to soft mud substrates, and crinoids, and developed at a more advanced stage of the transgression. Mixing of nearshore and offshore taxa may have resulted from winnowing of mud by storm waves, slow or nondeposition, and the effects of scavengers and burrowers moving through the substrate.

The absence of coastal marine facies and faunas (e.g. *Lingula* biofacies) at the base of the Winifrede Shale may be the result of the rapidity of the transgression or erosion of earlier marginal marine facies during coastal ravinement.

Facies 2

Facies 2 contains the deepest water facies within the Winifrede Shale, which is marked by a black, organic-rich band at the top. The distinctive, highly comminuted character of shell material in this facies is striking. Extensive fragmentation of shell material may arise from several mechanisms including turbulence-induced impact, sediment compaction, and biologic processes (shell boring and predation). Low sedimentation rates in quiet shelf settings expose skeletal material to the effects of predators and scavengers and may also lead to extensive disarticulation, fragmentation, and scattering of shell material (Dodd and Stanton, 1990).

The presence of shell fragments in alternating disseminated and concentrated modes could be explained by a periodic, winnowing of fragmented indigenous benthos. The well-sorted shell layers reflect periodic turbulence. However, the higher organic content toward the top of this interval suggests that dysaerobic conditions developed, inhibiting colonization by benthic shelled fauna. Alternatively, the shell fragments may have been transported into this environment from shallower, better-oxygenated substrates by seaward-directed, storm-generated back flow. This mechanism has been proposed to explain the origin of distal tempestites (Aigner, 1982).

In Pennsylvanian marine units from the North American Midcontinent, moderate-depth, stenotopic brachiopod-crinoid-rich biofacies are replaced by deeper water, phosphatic shales containing an oxygen-stressed molluscan benthos or fissile, black, phosphatic shales lacking indigenous benthos (i.e. "core shales" of Kansas cyclothems; Boardman and others, 1984).

A black shale that is similar to facies 2 has been described from the Devonian Hamilton Group of New York. Bedding plane concentrations of shell fragments alternating with laminae of sparsely fossiliferous, silty clay were interpreted to represent long periods of non-deposition between rapid deposition of mud tempestites in a low-oxygen environment (Parsons and others, 1988).

Oxygen-deficient bottom waters may result when bottom waters become insulated from well-oxygenated surface waters by strong density gradients associated with pycnoclines. These may result from stratification caused by vertical gradients in temperature or salinity. A rapid expansion of oxygen-deficient "puddles" of deep water occurs during the early stages of transgression in epicontinental seas. In areas of initially shallow water, these transgressive black shales rest on condensed, basal transgressive lags or unconformities (Wignall, 1991). Low oxygen may also be promoted by excess organic material from either terrestrial input or plankton blooms (Goldring, 1991).

The presence of *Planolites* and *Zoophycos*, which are traces of endobenthic deposit-feeders (Hantzschel, 1975; Pemberton and Frey, 1982), could indicate that at least some dissolved oxygen was intermittently present. These traces are known to occur in dysaerobic settings (Frey and Seilacher, 1980; Ekdale and others, 1984). Elevated levels of oxygen may have resulted from storm-generated bottom currents that produced the tempestites in this facies. Pyritization of burrows is also an indication of limited oxygen within the substrate, although this does not necessarily reflect conditions at the sediment-water interface. Alternatively, these burrows may represent later biogenic overprinting of sediments deposited under anoxic conditions.

Facies 3

The lower portion of facies 3 was deposited in an offshore, mud-dominated shelf setting. The accumulation of terrigenous mud in this type of setting is intermittent. The rate of deposition is highest immediately follow storms, and slowest during prolonged fair weather intervals (Johnson and Baldwin, 1986). Bottom currents may have reworked the mud substrate, episodically producing winnowed shelf condensates (distal tempestites) from soft ground benthos. Faunally barren shale intervals may have resulted from high turbidity and rapid mud deposition following storms, limited oxygen, or soupy substrates. Poorly sorted shell beds with articulated taxa represent colonization of the substrate during quiet, fair weather periods and minimal ter-

rigenous influx. The fauna was probably buried by more rapid mud deposition which minimized exposure and associated fragmentation and disarticulation.

Tempestites in facies 3 are thicker than those in facies 2, suggesting a more proximal origin (Aigner, 1982). Normal grading of brachiopods and crinoids in biomicrites at localities 2 and 3 is consistent with deposition under waning flow associated with storm-generated bottom currents.

The appearance and upward increase in abundance of siderite nodules in the upper part of facies 3 at locality 2 suggests that salinity began to decrease or fluctuate. Siderite nodules associated with marine shale are believed to form in settings with brackish water or fluctuating salinity combined with slightly acidic and oxygen deficient conditions (Woodland and Stenstrom, 1979; Maples, 1986).

Although the fossil assemblages of facies 1 and 3 are generally similar in that they contain abundant brachiopods and crinoids, important differences occur that may have paleoecologic significance. Facies 3 has little if any burrowing bivalves (such as *Astartella, Wilkingea*) that prefer siltier substrates, and contains spiny productids (*Desmoinesia, Antiquatonia*) and chonetids (*Rugosochonetes*) as dominant faunal elements. These brachiopods are well-adapted to soft clay-rich substrates (Rudwick, 1970).

The transition from offshore to nearshore may occur within facies 3 at the level where crinoids disappear and the assemblage becomes dominated by productids, chonetids, spiriferids, and compositids. Suspension-feeding echinoderms are sensitive to increased levels of turbidity; their water vascular systems are susceptible to clogging by suspended mud (Brett, 1998). The vertical persistence of brachiopods through facies 3 (localities 2 and 3) may reflect their higher tolerance for turbidity, afforded by their capacity to reject siliciclastic particles and their ability to "clam up" for brief periods during short-lived, high turbidity events (Rudwick, 1970). Modern brachiopods require full marine salinity and are intolerant of any dilution. Well-oxygenated water and low sedimentation rates are also requisite conditions for these sessile suspension feeders (Rudwick, 1965). The simultaneous disappearance of brachiopods and crinoids within facies 3 at locality 1 may have resulted from lowered salinity and/or possibly higher sedimentation rates.

Offshore Pennsylvanian assemblages are generally characterized by calcareous brachiopods, cephalopods, bivalves, pelmatozoans, bryozoans, and corals Williams, 1960; Ferm and Williams, 1965; Bretsky, 1969; Boardman and others, 1984; Gastaldo and others, 1989). The upper part of facies 3 closely resembles nearshore marine faunas and lithologies from the Allegheny Formation of Pennsylvania (Ferm and Williams, 1965). The number and types of brachiopod genera in this interval are also comparable to those found at an interpreted paleobathymetry of 6-15 m in the Minturn Formation of Colorado (Stevens, 1971).

Facies 4

The siderite bed may represent a single, widespread, sudden influx of freshwater into more saline waters. Laminated siderite beds are thought to form in this manner (Ferm, 1957; Woodland and Stenstrom, 1979). The occurrence of scattered plant detritus and increasing grain size within the overlying facies support the onset of more nearshore conditions. Rapid deposition and early diagenetic formation of siderite encapsulated the benthic assemblage and account for the high quality of preservation.

Facies 5

The coarser grain size, mica, and plant detritus of this interval all support a nearshore setting. Higher rates of sedimentation and freshwater influx are likely to have prevented colonization of the substrate by stenohaline invertebrates.

The alternation of very fine to fine sand layers with mud partings containing micas and finely divided plant detritus suggests that during much of the time, sediment accumulation occurred by deposition from suspension under waning flow conditions. These conditions characterize of the delta front where deltaic distributaries empty into marine waters. Here, lower density, freshwater discharge overrides the

denser seawater forming a buoyant surface plume. Flocculation of suspended load characterizes the resulting distal mouth bar deposits (Elliott, 1986).

During peak river discharge and high sediment load, traction transport due to density underflows may occur in the delta front environment (Wright and others, 1988; Bhattachyria and Walker, 1992). These processes are capable of producing scouring and filling like that which is occasionally evident in facies 5. Convolute bedding is consistent with rapid deposition and is commonly formed in delta front settings (Bhattachyria and Walker, 1992).

Geochemical Profiles and Paleosalinity

Certain trace elements have been shown to be sensitive to the salinity of the depositional environment. Boron content of the clay fraction of marine shales has been shown to be higher than for freshwater shales, using samples of similar mineralogy and diagenetic history (Walker, 1968, Harder, 1970; Couch, 1971). This is because the concentration of boron in seawater is 480 times higher than in fresh water (Ernst, 1970). The adsorption of boron onto clay particles requires prolonged exposure; boron content is not sensitive to rapid salinity changes (Hallam, 1981). Boron can be inherited, to some extent, indicating that provenance is another potential factor depending on climate and intensity of chemical weathering. In a humid climatic setting, like that interpreted for the Middle Pennsylvanian in the central Appalachian Basin (Cecil, 1990), substantial amounts of boron would have been leached. This reduces the likelihood that boron-rich clay was recycled from marine shales in the sediment source area.

Other trace elements that have been reported to be useful in distinguishing marine from freshwater clays include chromium, copper, gallium, nickel, and vanadium. However, these trace elements may be less reliable than boron (Shimp and others, 1969).

Vertical trends in these six trace elements through facies 1-5 (Table 1, Figures 3-5) suggest concentrations were influenced by conditions associated with different stages of the transgressive-regressive cycle. The facies interpretations and distribution of stenohaline taxa suggest that facies 1-3 accumulated under normal or full marine conditions and that facies 4 and 5 were likely to have experienced lower salinities. The results for boron, chromium, and nickel indicate a positive correlation with salinity. The average levels of these trace elements were higher at each location in facies 1-3 than in 4 and 5 (Table 2).

Boron levels generally are 50-100 ppm in facies 1-3 and 30-50 ppm in facies 4 and 5. An exception to these ranges occurs at location 2 in the upper part of facies 3 (145-175 cm). This stratigraphic interval consists of shale with abundant siderite nodules and is faunally barren. These attributes suggest that diluted salinities occurred in the upper part of facies 3 at this location. The average values for chromium and nickel are also lower for this interval, further suggesting lowered salinity.

Factors other than salinity could be responsible for the differences in boron levels between facies 1-3 and 4-5. Higher rates of deposition can accompany regression particularly if prograding deltaic facies are involved. This would decrease the exposure time of clays to seawater (Harder, 1961; Porrenga, 1967). Porrenga (1967) analyzed boron levels in the clay fraction of 150 samples from the Niger Delta, and observed that concentrations (in ppm) were lowest for river clays from freshwater environments(< 50; average: 40), intermediate for estuarine and nearshore clays from brackish environments (50-75; average: 60), and highest for normal marine environments (>75; average: 110). Boron levels in marine samples were proportional to water depth and to distance from shore. This was attributed to a gradual change in clay mineralogy involving less kaolinite and more montmorillonite seaward, and higher residence (exposure) time on the seafloor due to lower rates of sedimentation and greater transport distance.

Degens and others (1957) analyzed trace elements in the clay fraction of Middle Pennsylvanian shales from the Appalachian Basin. Shales from the Alleghenny Formation of Pennsylvania were interpreted as freshwater, brackish,

and marine based on faunal content. Boron levels were highest in marine shales. Gallium levels were highest in freshwater shales. Nickel and vanadium were found to be concentrated in the organic fraction of marine shales, whereas copper was more abundant in the organic matter of freshwater shales.

Mean values for nickel were consistently higher in facies 1-3 than 4-5, with vanadium following the same pattern except at locality 2. These trends for nickel and vanadium may be due to a greater proportion of terrestrial plant detritus in the nearshore/delta front facies.

Biofacies of the Magoffin and Winifrede Shales

Magoffin Shale biofacies and their paleoecologic implications have been addressed by Dennis and Lawrence (1979) and by Bennington (1996, 1998).

Dennis and Lawrence distinguished a brachiopod-dominated assemblage interpreted to be a nearshore biofacies, and a mollusk-dominated assemblage which represented a deeper water biofacies. Bennington (1996) distinguished the following biofacies in northeastern Kentucky in ascending order: (1) brachiopod, (2) gastropod/*Pleurophorella,* (3) *Posidonia,* (4) gastropod/nuculoid bivalve, (5) brachiopod. This sequence of biofacies was interpreted as a series of depth-related faunal associations that developed during a transgressive-regressive cycle. The midpoint of the cycle is represented by the *Posidonia*-bearing, fissile black shale which was attributed to anoxia developed beneath a pycnocline at the peak of the transgression.

Bennington (1996) analyzed one outcrop (his M74) in the Winifrede Shale type area. Here, he recognized lower and upper brachiopod-dominated biofacies corresponding to facies 1 and 3 of this study, but was unable to recognize either of his deeper water biofacies in the intervening interval (facies 2 of this study). This could be the result of the condensed character of the Winifrede Shale in its type area along the east-west-trending hinge of the basin. The rest of Bennington's outcrops were located further south where subsidence rates and accommodation space were greater. Nuculoids such as *Astartella* and gastropods generally are common in facies 1 of this study. As stated previously, this basal interval may represent mixing of nearshore and offshore faunal elements.

The vertical distribution of proximal and distal tempestites and the presence of a carbonaceous shale band at all 3 localities of this study support the idea that brachiopod-dominated facies 1 and 3 bracket a deeper water, dysaerobic facies in the Winifrede Shale type area.

Sequence Stratigraphy

The Magoffin Shale has been interpreted as part of a fourth order sequence (Aitken and Flint, 1994; Bennington, 1998). In this and other marine shales in the Breathitt Group, Aitken and Flint typically placed the boundary between the transgressive and highstand systems tracts (TST, HST) at the top of marine shelf facies association. This shelf facies was distinguished by gray to black, laminated and bioturbated siltstone with marine bivalves, crinoids, brachiopods, cephalopods, trace fossils and plant fragments.

Bennington (1996, 1998) was able to distinguish within what Aitken and Flint (1994) treated as a transgressive systems tract a series of depth-related biofacies. He concluded that Magoffin seaway became sufficiently deep to develop a pycnocline. This resulted in a progressive sequence from oxygenated to dysaerobic to anoxic benthic environments during transgression which was mirrored in reverse during regression. A critical difference in Bennington's (1998) treatment of systems tracts was his placement of the top of TSTs at the top of a carbonaceous shale *within* the marine shelf facies. This approach more accurately defines the position of the maximum flooding surface and is followed here. Much of the strata that Aitken and Flint (1994) treated as TST are actually within the lower portion of the HST.

The base of the transgressive systems tract is defined by the marine-flooding surface. This surface separates vertically adjacent facies that reflect a sharp increase in water depth. The deepening is often accompanied by minor sub-

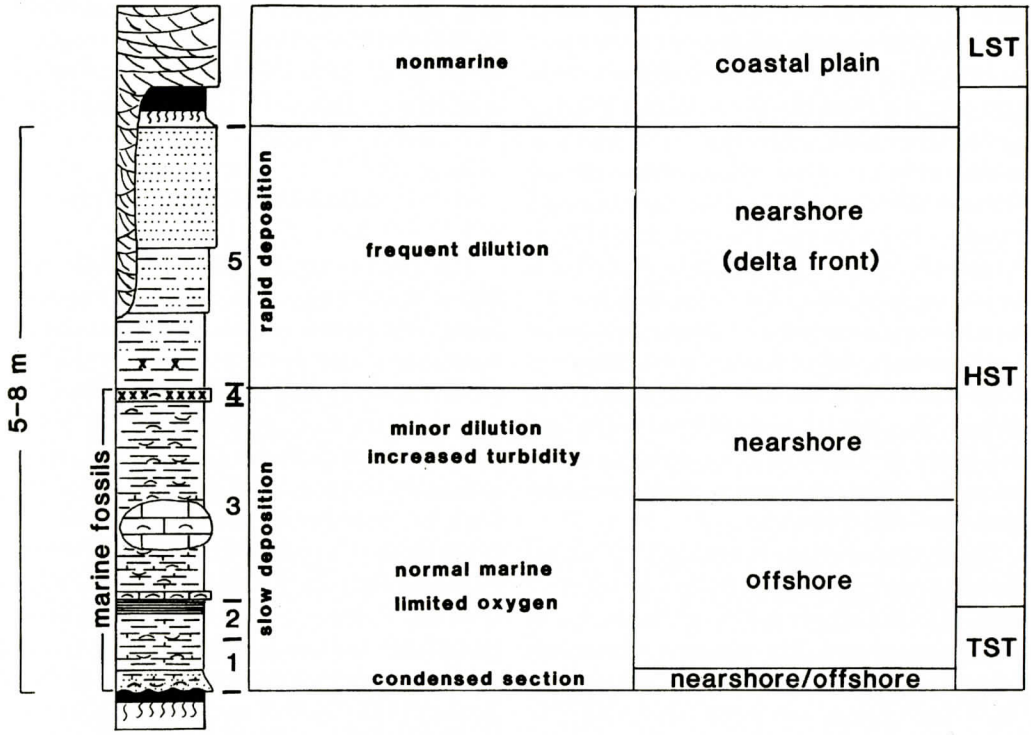

Figure 8. Generalized stratigraphic column through Winifrede Shale with interpreted paleoenvironments and systems tracts (modified from Martino, 1996).

marine erosion and nondeposition and a minor hiatus may develop (Van Wagoner and others, 1988). In this study, the marine flooding surface corresponds to the contact between facies 1 of the Winifrede Shale and underlying coal or underclay. This basal mudstone may represent a condensed section, an interval of thin marine deposits formed by very slow deposition during regional transgression of the shoreline (Van Wagoner and others, 1988). The condensed section usually forms in the TST or distal HST. The top of the TST corresponds to the maximum water depth which is represented by the carbonaceous shale at the top of facies 2 (Figure 8).

The HST includes facies 3-5, i.e. the calcareous shale interval above the carbonaceous shale biomicrite, siderite bed, and coarsening-upward sequence. The truncation of the section at locality 3 by alluvial channel facies may be associated with a type 1 sequence boundary related to the next sea level lowstand (Figure 8). During the initial phase of the highstand, terrigenous sediments were trapped in coastal estuarine sediment sinks. Very slow deposition maximized the degree of degradation of skeletal material by biologic and hydraulic processes. During this time terrigenous influx was sufficiently limited to allow for the development of stenotopic benthic marine faunas. As estuaries filled, coastal progradation occurred which initially caused an increase in the rate of mud deposition in shelf areas. This resulted in gradual improvement in preservation of shelf faunas. With increasing proximity to the shoreline, rapid depositional episodes were accompanied by freshwater influx which promoted the development of siderite. Ultimately, the frequent reductions in salinity and high rates of deposition prevented benthos from colonizing nearshore and coastal facies.

CONCLUSIONS

The Winifrede Shale in its type area contains

sedimentary facies that comprise transgressive and highstand systems tracts of a fourth-order stratigraphic sequence. A ravinement surface is preserved at its base that is overlain by a highly fossiliferous sandy mudstone. The fauna is dominated by calcareous brachiopods, and also contains bivalves, gastropods, cephalopods, crinoids, and bryozoans. The basal facies (1 and 2) represent a condensed section formed at a time when sediment influx to the shelf was reduced during transgression. Some mixing of nearshore and offshore faunal elements appears to have occurred at the base of this trangressive interval. Maximum transgression was accompanied by the development of a pycnocline and is marked by a thin carbonaceous shale containing distal bioclastic tempestites.

The highstand systems tract begins above the carbonaceous shale and contains an offshore fauna in its lower part that is similar to that of the basal mudstone. The presence of proximal tempestites indicates the water depth was between fair weather and storm wave base. The disappearance of nearly all mollusks, crinoids, and bryozoans in the upper part of the calcareous shale, and increase in siderite reflect the onset of regression and more turbid nearshore conditions. Bedded siderite may have formed during periodically reduced salinity. The marine invertebrate-bearing interval is capped by a faunally barren, coarsening-upward shale and sandstone that was rapidly deposited in a delta front.

Trace element profiles from the clay fraction of the Winifrede Shale show trends that may broadly reflect salinity and depositional conditions. The average levels of boron are higher in the transgressive and early highstand systems tract as would be expected for normal marine salinity and slow sedimentation rates. During late highstand as deltaic regression progressed, more rapid deposition and dilution led to lower boron levels. Nickel and vanadium also show generally higher values in more fully marine facies which may be due to their tendency to become concentrated in marine organic carbon.

The absence of marine fossils in black shales of suspected marine origin elsewhere in the Kanawha Formation has been attributed to anoxia, high turbidity, or frequent dilution (Martino, 1994, 1996). The depositional origin of these faunally barren shales may be clarified by trace element analyses which could help reveal paleosalinity conditions.

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REFERENCES

Aigner, T., 1982, Calcareous tempestites: storm-dominated stratification Upper Muschelkalk limestones (Middle Triassic, SW Germany), *in* Einsele, G. and Seilacher, A., eds., Cyclic and Event Stratification: Springer-Verlag, Berlin, p. 180-198.

Aitken, J. F., and Flint, S. S., 1994, High-frequency sequences and the nature of incised-valley fills in fluvial systems of the Breathitt Group (Pennsylvanian), Appalachian foreland basin, eastern Kentucky, *in* Dalrymple, R. W., Boyd, R., and Zaitlin, B. A., eds., Incised-valley Systems: Origin and Sedimentary Sequences: Society of Economic Paleontologists and Mineralogists Special Publication No. 51, p. 353-368.

Bennington, J. B., 1996, Stratigraphic and biofacies patterns in the Middle Pennsylvanian Magoffin Marine Facies in the Appalachian Basin, U.S.A.: International Journal of

Coal Geology, v. 31, p. 169-193.

Bennington, J. B., 1998, Biofacies as indicators of a forced regression in the Middle Pennsylvanian Magoffin marine facies, Appalachian Basin, U.S.A: Geological Society of America, Abstracts with Program, v. 30, no. 2, p. 5.

Bhattacharya, J. P., and Walker, R. G., 1992, Deltas: *in* Walker, R. G., and James, N. P., eds., Facies Models – Response to Sea Level Change: Geological Association of Canada, p. 157-177.

Blake, B. M. Jr., Keiser, A. F., and Rice, C. L., 1994. Revised stratigraphy and nomenclature for the Middle Pennsylvanian Kanawha Formation in southern West Virginia, *in* Rice, C. L., ed., Elements of Pennsylvanian Stratigraphy, Central Appalachian Basin: Geological Society of America Special Paper 294, p. 41-53.

Blake, B. M. Jr., 1998, Regional stratigraphy and coal geology of the Kanawha Formation in southern West Virginia, *in* Martino, R. L., Blake, B. M., Jr., Grady, W. C., and Eble, C. F., Coal Geology, Paleobotany, and Regional Stratigraphy of the Middle Part of the Kanawha Formation, Southern West Virginia: Geological Society of America, Southeastern Section, Coal Division Field Trip Guidebook, p. 1-14.

Boardman, D. R., II, Mapes, R. H., Yancey, T. E., and Malinky, J. M., 1984, A new model for depth-related allogenic community succession within North

American Pennsylvanian cyclothems and implications on the black shale problem, *in* Hyne, N. J., ed., Limestones of the Mid-Continent: Tulsa Geological Society Special Publication 2, p. 141-182.

Bretsky, P. W., 1969, Evolution of Paleozoic benthonic marine invertebrate communities: Paleogeography, Paleoclimatology, Paleoecology, v. 6, p. 45-59.

Brett, C. E., 1998, Sequence stratigraphy, paleoecology, and evolution: biotic clues and responses to sea-level fluctuations: Palaios v. 13, p. 241-262.

Cecil, C. B., 1990, Paleoclimate controls on stratigraphic repetition of chemical and siliciclastic rocks: Geology, v. 18, p. 533-536.

Couch, E. L., 1971, Calculation of paleosalinities from boron and clay mineral data: American Association of Petroleum Geologists Bulletin, v. 55, p. 1829-1837.

Degens, E. T., Williams, E. G., and Keith, M. L., 1957, Environmental studies of Carboniferous sediments. Part I: Geochemical criteria for differentiating marine from freshwater shales: American Association of Petroleum Geologists Bulletin, v. 41, p. 2427-2455.

Dennis, A. M., and Lawrence, D. R., 1979, Paleoecology in the Magoffin zone, Pennsylvanian Breathitt Formation near Hazard, Kentucky, *in* Ferm, J. C., Horne, J. C., Weisenfluh, G. A., Staub, J. R., eds., Carboniferous Depositional Environments in the Appalachian Region: Carolina Coal Group, Department of Geology, University of South Carolina, Columbia, S. C., p. 256-258.

Dodd, J. R., and Stanton, R. J., Jr., 1990, Paleoecology, Concepts and Applications: Second Edition, John Wiley and Sons, New York, 502 p.

Ekdale, A. A., Bromley, R. G., and Pemberton, S. G., 1984, Ichnology---Trace Fossils in Sedimentology and Stratigraphy: Society of Economic Paleontologists and Mineralogists, Short Course No. 15, 317 p.

Elliott, T., 1986, Deltas, *in* Reading, H. G., ed., Sedimentary Environments and Facies: Blackwell Scientific Publications, Oxford, p. 113-154.

Ernst, W., 1970, Geochemical Facies Analysis: Elsevier, Amsterdam, 152 p.

Ferm, J. C., 1957, Petrology of the Kittanning Formation near Brookville, Pennsylvania: Unpublished Ph. D. dissertation, Pennsylvania State University, 381 p.

Ferm, J. C., and Williams, E. G., 1965, Characteristics of a Carboniferous marine invasion in western Pennsylvania: Journal of Sedimentary Petrology, v. 35, p. 319-330.

Frey, R. W., and Seilacher, A., 1980, Uniformity in marine invertebrate ichnology: Lethaia, v. 13, p. 183-207.

Fursich, F. T., 1978, The influence of faunal condensation and mixing on preservation of fossil benthic communities: Lethaia, v. 11, p. 243-250.

Gastaldo, R. A., Gibson, M. A., and Gray, T. D., 1989, An Appalachian-sourced deltaic sequence, northern Alabama, U. S. A.: biofacies-lithofacies relationships and interpreted community patterns: International Journal of Coal Geology, v. 12, p. 225-257.

Gastaldo, R. A., Demko, T. A., and Lui, Y., 1991, Condensed sections and the regressive character of Pennsylvanian marine facies in the Black Warrior Basin: Geological Society of America, Abstracts with Programs, v. 23, no. 1., p. 33.

Gillespie, W. H. and Pfefferkorn, H. H., 1979, Distribution of commonly occurring plant megafossils in the proposed Pennsylvanian System stratotype, *in* Englund, K. J., Arndt, H. H., and Henry, T. W., eds., Proposed Pennsylvanian System Stratotype, Virginia and West Virginia: American Geologic Institute Selected Guidebook Series 1., p. 87-96.

Goldring, R., 1991, Fossils in the Field, Information Potential and Analysis: John Wiley and Sons, New York, 218 p.

Harder, H., 1961, Contribution to the geochemistry of Boron: II. Boron in sediments, *in* Walker, C. T., ed., Geochemistry of Boron: 1975, Benchmark Papers in Geology, v. 23, Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pa., p. 83-104

Harder, H., 1970, Boron content of sediments as a tool in facies analysis: Sedimentary Geology, v. 4, p. 153-175.

Hallam, A., 1981, Facies Interpretation and the Stratigraphic Record: W. H. Freeman, San Francisco, 291 p.

Hantzschel, W., 1975, Trace Fossils and Problematica, *in* Teichert, C., ed., Treatise of Invertebrate Paleontology, Part W, Miscellanea Supplement 1: Geological Society of America and University of Kansas Press, Lawrence, Kansas, 269 p.

Henry, T. W., and Gordon, M., Jr., 1979, Late Devonian through Early Permian(?) invertebrate faunas in proposed Pennsylvanian System stratotype area, *in*

Englund, K. J., Arndt, H. H., and Henry, T. W., eds., Proposed Pennsylvanian System Stratotype, Virginia and West Virginia: American Geologic Institute Selected Guidebook Series 1., p. 97-104.

Jackson, J. A., 1997, Glossary of Geology: Fourth edition, American Geologic Institute, Alexandria, Virginia, 769 p.

Johnson, H. D., and Baldwin, C. T., 1986, Shallow siliciclastic seas: *in* Reading, H. G., ed., Sedimentary Environments and Facies: Blackwell Scientific Publications, Oxford, p. 229-282.

Lui, Y., and Gastaldo, R. A., 1989, Characteristics and implications of a ravinement surface in the upper Pottsville Formation, northwestern Alabama: Geological Society of America, Abstracts with Programs, v. 21, no. 3, p. 48.

Maples, C. G., 1986, Enhanced paleoecological and paleoenvironmental interpretations result from analysis of early diagenetic concretions in Pennsylvanian shales: Palaios, v. 1, p. 512-516.

Martino, R. L., 1994, Facies analysis of Middle Pennsylvanian marine units, southern West Virginia, *in* Rice, C. L., ed., Elements of Pennsylvanian Stratigraphy, Central Appalachian Basin: Geological Society of America Special Paper 294, p. 69-86.

Martino, R. L. 1996, Stratigraphy and depositional environments of the Kanawha Formation (Middle Pennsylvanian), southern West Virginia, U. S. A.: International Journal of Coal Geology, v. 31, p. 217-248.

Outerbridge, W. F., 1976, The Magoffin Member of the Breathitt Formation: *in* Cohee, G. V., and Wright, W. B., eds., Changes in Stratigraphic Nomenclature in the U. S. Geological Survey, 1975: U. S. Geological Survey Bulletin 1422-A, p. A64-A65.

Parsons, K. M., Brett, C. E., and Miller, K. B., 1988, Taphonomy and depositional dynamics of Devonian shell-rich mudstones: Paleogeography, Paleoclimatology, Paleoecology, v. 63, p. 109-139.

Pemberton, S. G., and Frey, R. W., 1982, Trace fossil nomenclature and the *Planolites-Paleophycus* dilemma: Journal of Paleontology, v. 56, p. 843-881.

Porrenga, D. H., 1967, Clay mineralogy and geochemistry of recent marine sediments in tropical areas, *in* Walker, C. T., ed., Geochemistry of Boron: 1975, Benchmark Papers in Geology v. 23, Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pa., p. 283-306.

Price, W. A., 1914, Notes on the paleontology of Kanawha County, *in* Krebs, C. E., and Teets, D. D., Jr., Kanawha County: West Virginia Geological Survey County Report, p. 639-655.

Riley, N., and Turner, N., 1994, The correlation of mid-Westphalian marine bands between the central Appalachian basin (USA) and the United Kingdom: XIII International Congress on Carboniferous-Permian, Abstracts, p. 122.

Rudwick, M. J. S., 1965, Ecology and Paleoecology: *in* Moore, R. C., ed., Treatise on Invertebrate Paleontology, Part H, Brachiopoda, v. 1, University of Kansas Press and Geological Society of America, Lawrence, Kansas, p. H199-H214.

Rudwick, M. J. S., 1970, Living and Fossil Brachiopods: Hutchinson University Library, London, 199 p.

Shimp, N. F., Witters, J., Potter, P. E., and Schleicher, J. A., 1969, Distinguishing marine and freshwater muds: Journal of Geology, v. 77, p. 566-580.

Speyer, S. E., and Brett, C. E., 1988, Taphofacies models for epeiric sea environments: Middle Paleozoic examples: Paleogeography, Paleoclimatology, Paleoecology, v. 63, p. 225-262.

Stevens, C. H., 1971, Distribution and diversity of Pennsylvanian marine faunas relative to water depth and distance from shore: Lethaia, v. 4, p. 403-412.

Swift, D. J. P., 1968, Coastal erosion and transgressive stratigraphy: Journal of Geology, v. 76, p. 444-456.

Van Wagoner, J. C., Posamentier, H. W., Mitchum, R. M., Vail, P. R., Sarg, J. F., Loutit, T. S., and Hardenbol, J., 1988, An overview of the fundamentals of sequence stratigraphy and key definitions, *in* Wilgus, C. K., Hastings, B. S., Kendall, C. G. St. C., Posamentier, H. W., Ross, C. A., and Van Wagoner, J. C., eds., Sea-level Changes: an Integrated Approach: Society of Economic Paleontologists and Mineralogists Special Publication No. 42, p. 39-45.

Walker, C. T., 1968, Evaluation of boron as a paleosalinity indicator and its application to offshore prospects: American Association of Petroleum Geologists Bulletin, v. 52, p. 751-766.

Wagner, R. H., and Winkler Prins, C. F., 1991, Major subdivisions of the Carboniferous System: Compte Rendu, 11th Congres International de Geologie du Carbonifere (Bejing, 1987), I, p. 213-245.

Wanless, H. R., 1939, Pennsylvanian correlations in the Eastern Interior and Appalachian coal fields: Geological Society of America Special Paper 17, 130 p.

Wignall, P. B., 1991, Model for transgressive black shales?: Geology, v. 19, p. 167-170.

Williams, E. G., 1960, Marine and fresh water fossiliferous beds in the Pottsville and Allegheny Groups of western Pennsylvania: Journal of Paleontology, v. 34, p. 908-922.

Woodland, B. G., and Stenstrom, R. C., 1979, The occurrence and origin of siderite concretions in the Francis Creek Shale (Pennsylvanian) of northeastern Illinois, *in* Nitecki, M. H., ed., Mazon Creek Fossils: Academic Press, New York, p. 69-103.

Wright, L. D., Wiseman, W. J., Bornhold, B. D., Prior, D. B., Suhayda, J. N., Keller, G. H., Yang, Z. S., and Fan, Y. B., 1988, Marine dispersal and deposition of Yellow River silts by gravity-driven underflows: Nature, v. 332, p. 629-632.

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NEW MATERIAL OF *EOCETUS WARDII* (MAMMALIA, CETACEA), FROM THE MIDDLE EOCENE OF NORTH CAROLINA

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ABSTRACT

New material of the protocetid archaeocete, *Eocetus wardii*, from the late Lutetian (middle Eocene, 43-44 Ma) Castle Hayne Formation, North Carolina shows that *E. wardii* had 12 thoracic vertebrae, a scapula like those of later-occurring basilosaurid archaeocetes, and an unusual sternum unlike those of basilosaurids. This material includes sixteen vertebrae (C7 to L2 plus a caudal), the right scapula, both stylohyals, the manubrium and a mesosternal element of the sternum, seven complete ribs and various rib fragments. The broad, fan-shaped scapula indicates that *Eocetus* possessed a proximal forelimb like those of later, fully aquatic basilosaurids, which, along with the potentially weight-bearing innominate (pelvis exclusive of the sacral vertebrae) suggests that *Eocetus wardii* may have been capable of both terrestrial and aquatic locomotion.

INTRODUCTION

Over 25 specimens of archaeocete cetaceans have been discovered in the Eocene rocks of North Carolina and deposited in the North Carolina Museum of Natural Sciences. Most of these specimens were previously identified as either *Basilosaurus cetoides* or *Zygorhiza kochii*. While both taxa appear to have been present in North Carolina during the Eocene, many of the specimens in the collection are actually *Eocetus wardii*, a recently described protocetid. These specimens have greatly expanded the knowledge of *E. wardii*, and are described for the first time here.

The specimens discussed here are located at the North Carolina Museum of Natural Sciences, Raleigh, North Carolina (NCSM); Staatliches Museum für Naturkunde, Stuttgart (SMNS); and the United States National Museum of Natural History, Washington DC (USNM).

SYSTEMATIC PALEONTOLOGY

Order Cetacea Brisson, 1762
Suborder Archaeoceti Flower, 1883
Family Protocetidae Stromer, 1908

Type Genus. — *Protocetus* Fraas, 1904.

Diagnosis. — Thewissen and others (1996) characterized members of the family Protocetidae as having a supraorbital shield, greatly enlarged mandibular foramina, a reduced nasopharyngeal duct (compared to *Ambulocetus*), and reduced hind limbs. All non-basilosaurid archaeocetes, including protocetids, are characterized by the retention of M^3, lack of well-developed accessory denticles on the cheek teeth, and presence of innominata that lack a greatly elongated pubis, and possess a craniodorsally oriented ilium (rather than caudodorsally oriented as in Basilosauridae). The family Protocetidae is most likely paraphyletic, giving rise to the Basilosauridae (Thewissen and others, 1996; see below).

Included Subfamilies. — Protocetinae Stromer, 1908; Indocetinae Gingerich and others, 1993.

Occurrence. — Protocetines are known from Indo-Pakistan, Egypt, west Africa, and eastern North America and range from late early Lutetian to early Bartonian in age. Indocetines (sensu Uhen, 1999) are known only from the Lutetian of Indo-Pakistan (Sahni and Mishra, 1975; Gingerich and others, 1995)

Discussion. — Many species of protocetids

from North America are currently being described and others await description (Geisler and others, 1996; McLeod and Barnes, 1996; personal observation). These new species indicate a high diversity of protocetids in North America by the Bartonian.

Subfamily Protocetinae Stromer, 1908

Type Genus. — *Protocetus* Fraas, 1904.

Diagnosis. — Members of the subfamily Protocetinae have either a single sacral vertebra connected to the innominata or they have no sacral vertebrae that are directly connected to the innominata or connected to one another. See Uhen (1999) for a discussion of the differences between and composition of the subfamilies of Protocetidae, the Protocetinae and Indocetinae.

Included Genera. — *Eocetus* Fraas, 1904; *Pappocetus* Andrews, 1920; *Babiacetus* Trivedy and Satsangi, 1984; *Takracetus* Gingerich, Arif, and Clyde, 1995; *Gaviacetus* Gingerich, Arif, and Clyde, 1995; *Georgiacetus* Hulbert and others, 1998, *Natchitochia* Uhen, 1998a.

Occurrence. — Domanda Formation of Pakistan, early Lutetian (Gingerich and others, 1995); Drazinda Formation of Pakistan, late Lutetian (Gingerich and others, 1995); "gypseous shale bed of Babia Hills", India (Trivedy and Satsangi, 1984); Giushi Formation of Egypt, Bartonian (Gingerich, 1992); Mokattam Formation of Egypt, middle Lutetian (Gingerich, 1992); Ameki Formation of Nigeria, middle Eocene (Halstead and Middleton, 1974); Castle Hayne Formation of North Carolina, late Lutetian (Uhen, 1999); upper Santee Formation of South Carolina, latest Lutetian or earliest Bartonian (Albright, 1996); Cross Formation of South Carolina, early Bartonian (Geisler and others, 1996); "Blue Bluff" unit of the McBean Formation of Georgia, latest Lutetian or earliest Bartonian (Hulbert and others, 1998); Cook Mountain Formation of Louisiana, early Bartonian (Uhen, 1998a).

Discussion. — The subfamily Protocetinae is most likely paraphyletic, giving rise to the Basilosauridae (Thewissen and others, 1996; see Phylogenetic Relationships below). All described New World protocetids belong to the more derived Protocetinae, while indocetines appear to have been restricted to the Old World. Additional specimens of individual vertebrae of presumed protocetids have been found in Texas (Kellogg, 1936), and Alabama (Uhen, personal observation).

Genus *Eocetus* Fraas, 1904

Type Species. — *Eocetus schweinfurthi* Fraas, 1904.

Diagnosis. — *Eocetus* is larger than all other protocetines for which the skull is known (*Protocetus*, *Gaviacetus*, *Takracetus*, *Babiacetus*, and *Georgiacetus*). Based on the size of their vertebrae, *Pappocetus* and *Natchitochia* could rival *Eocetus* in skull length, but both have vertebrae assigned to them which are very different from those of *Eocetus* (Halstead and Middleton, 1974; Uhen, 1998b). The vertebrae of *Eocetus* have a distinctive pock-marked texture. The pock marks are actually small vascular channels that penetrate deeply into the bone. In addition, the centra, neural arches, neural spines, and transverse processes of the lumbar vertebrae are antero-posteriorly elongate, unlike the vertebrae of *Pappocetus* and *Natchitochia*.

Included species. — *Eocetus wardii* (Uhen, 1999)

Occurrence. — *Eocetus* is known from the Giushi Formation, Gebel Mokattam, Egypt, which is early Bartonian in age (Gingerich, 1992) and the late middle Eocene (latest Lutetian, NP 16) Castle Hayne Formation of North Carolina (Uhen, 1999).

Discussion. — *Eocetus* in North America is apparently restricted to North Carolina despite intensive collecting of other protocetid archaeocetes of middle Eocene age in nearby South Carolina.

Eocetus schweinfurthi Fraas, 1904

Type Specimen. — SMNS 10986, damaged skull. The isolated teeth illustrated by Fraas (1904, plate 2, figure 10-11) are missing.

Diagnosis. — *Eocetus schweinfurthi* differs from *Eocetus wardii* in having a dorsoventrally thinner rostrum and having laterally flaring in-

Table 1. Measurements of vertebrae from *Eocetus wardii*, NCSM 11284.

	measurement (MM)						
vertebra	ventral length	anterior width	anterior height	dorsal length	posterior width	posterior height	epiphyses
C7	32	56.4	48.3		71	47	missing both epiphyses
T1	51	62.3	41.5		70	42	missing posterior epiphysis
T2	63.3	59.4	45.2		64.3	47.5	anterior epiphysis loose
T3	63.7	67.1	52		69	50.8	missing anterior epiphysis
T4	60.3	64.2	48	56.3	69.6	48	missing anterior epiphysis
T5	58.4	68	50.5		74.5	50.5	missing anterior epiphysis
T6	74.3	66.8	56		79.4	54.1	missing neither epiphysis
T7	74	72.2	56.7		86.6	59.4	missing anterior epiphysis
T8	78.2	73	60.6		93.6	62.7	missing posterior epiphysis
T9	80.5	84.2	60.2		98.8	65.9	missing both epiphyses
T10	89.5	93	67.9		108.6	73.5	missing both epiphyses
T11	111	85.9	72.5		108.5	80.7	missing posterior epiphysis
T12	118.1	97	81.8		111.9	86.4	missing both epiphyses
L1	134	95.9	84.1		118.8	89.3	missing both epiphyses
L2	135.1	104.1	86.7		120.3	94.5	missing both epiphyses

cisor alveoli, which *E. wardii* lacks. The embrasure pits located posterior to I^3 and C^1 on the premaxillae are also more deeply incised in *E. schweinfurthi* than they are in *E. wardii*.

Referred Specimens. — NSFM 4470, two lumbar vertebrae.

Type Locality. — Gebel Mokattam, near Cairo, Egypt. See Gingerich (1992) for an extensive discussion of the locality and its age.

Occurrence. — Giushi Formation, early Bartonian, Gebel Mokattam, Egypt (Gingerich, 1992).

Description. — See Fraas (1904) and Uhen (1999) for descriptions of *Eocetus schweinfurthi*.

Discussion. — At least one tooth remains embedded in matrix on the base of the skull of the type specimen of *Eocetus schweinfurthi*. Further preparation may reveal even more teeth for comparison with other protocetids.

Eocetus wardii Uhen, 1999

Type Specimen. — USNM 310633, thoracic, lumbar, and caudal vertebrae, ribs, portions of skull, and right innominate.

Diagnosis. — See the diagnosis for *Eocetus wardii* for characteristics that differentiate it from *Eocetus schweinfurthi* within the genus *Eocetus*.

Referred Specimens. — USNM 449548, thoracic vertebra and portions of two ribs; USNM 449549, centrum of a lumbar vertebra; NCSM 11284, stylohyals, sixteen vertebrae (C7-L2, Ca?), manubrium and mesosternum, ribs, and right scapula; NCSM 11297, one thoracic and three lumbar vertebrae; NCSM 12531, supraoccipital; NCSM 13513, transverse process of a lumbar vertebra; NCSM 13514, vertebral body (C7 or T1?); NCSM 13434, lumbar vertebral body; NCSM 13676, proximal end of rib; NCSM 13678, partial caudal(?) vertebra; NCSM 15663, partial manubrium.

Type Locality. — Lanier's Pit, Maple Hill, Pender County, North Carolina. At the end of a dirt road, extending southeastward from SSR 1532, about 0.6 km southwest from the junction of SSR 1532 and NC 50, 3.5 km southeast from Maple Hill (Feldmann and others, 1998). Lanier's Pit is located at approximately 77° 40' 30" W longitude, 34° 37' 30" N latitude, USGS Maple Hill quadrangle, 7.5' series.

Occurrence. — All known specimens are from the Castle Hayne Formation of North Carolina. The Comfort Member, where the type specimen was found (Uhen, 1999), has been shown to be latest Lutetian (NP 16) based on both micro- and macrofossil correlations (Hazel and others, 1984), which would place it in sequence stratigraphic cycle TA 3.5 (Harris and Laws, 1997), equivalent to local Castle Hayne sequence 2

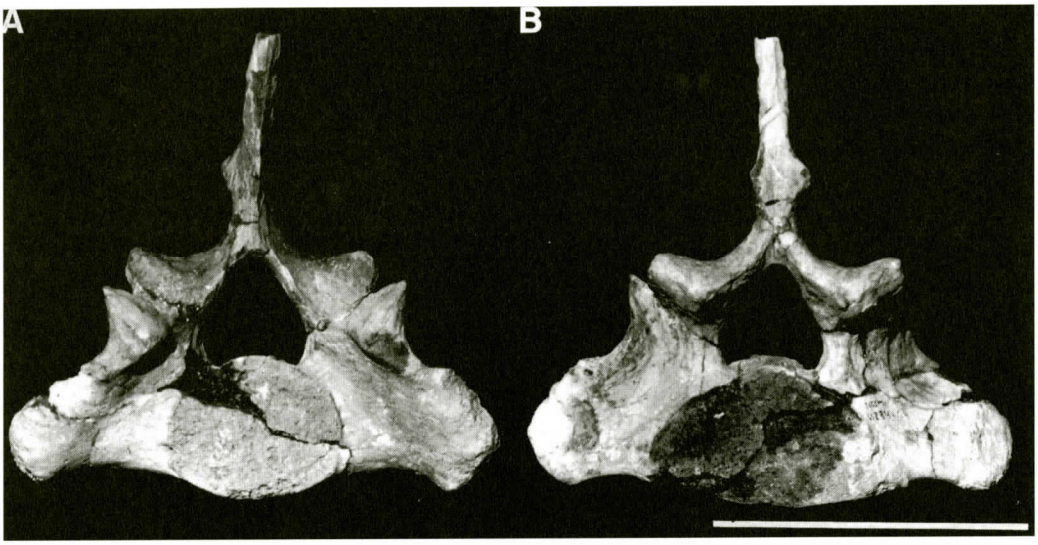

Figure 1. C7 of *Eocetus wardii*, NCSM 11284. A, anterior view; B, posterior view. Scale bar is 10 centimeters.

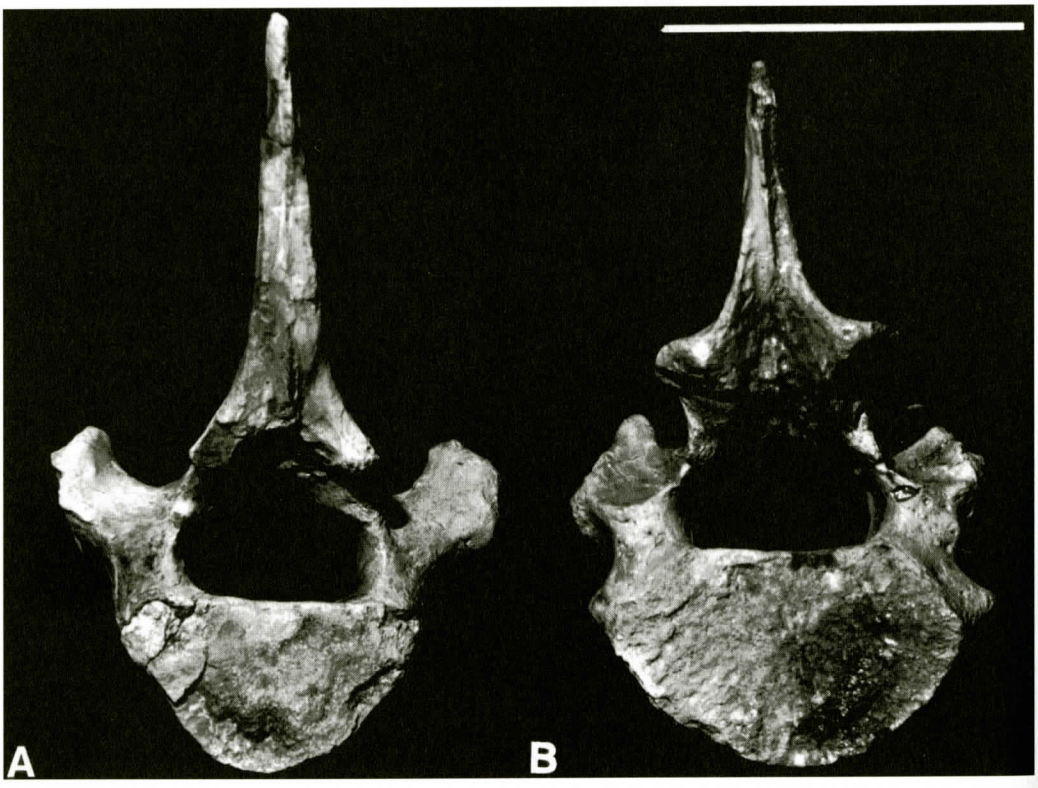

Figure 2. T2 and T5 of *Eocetus wardii* in posterior view, NCSM 11284. A, T3; B, T5. Scale bar is 10 centimeters.

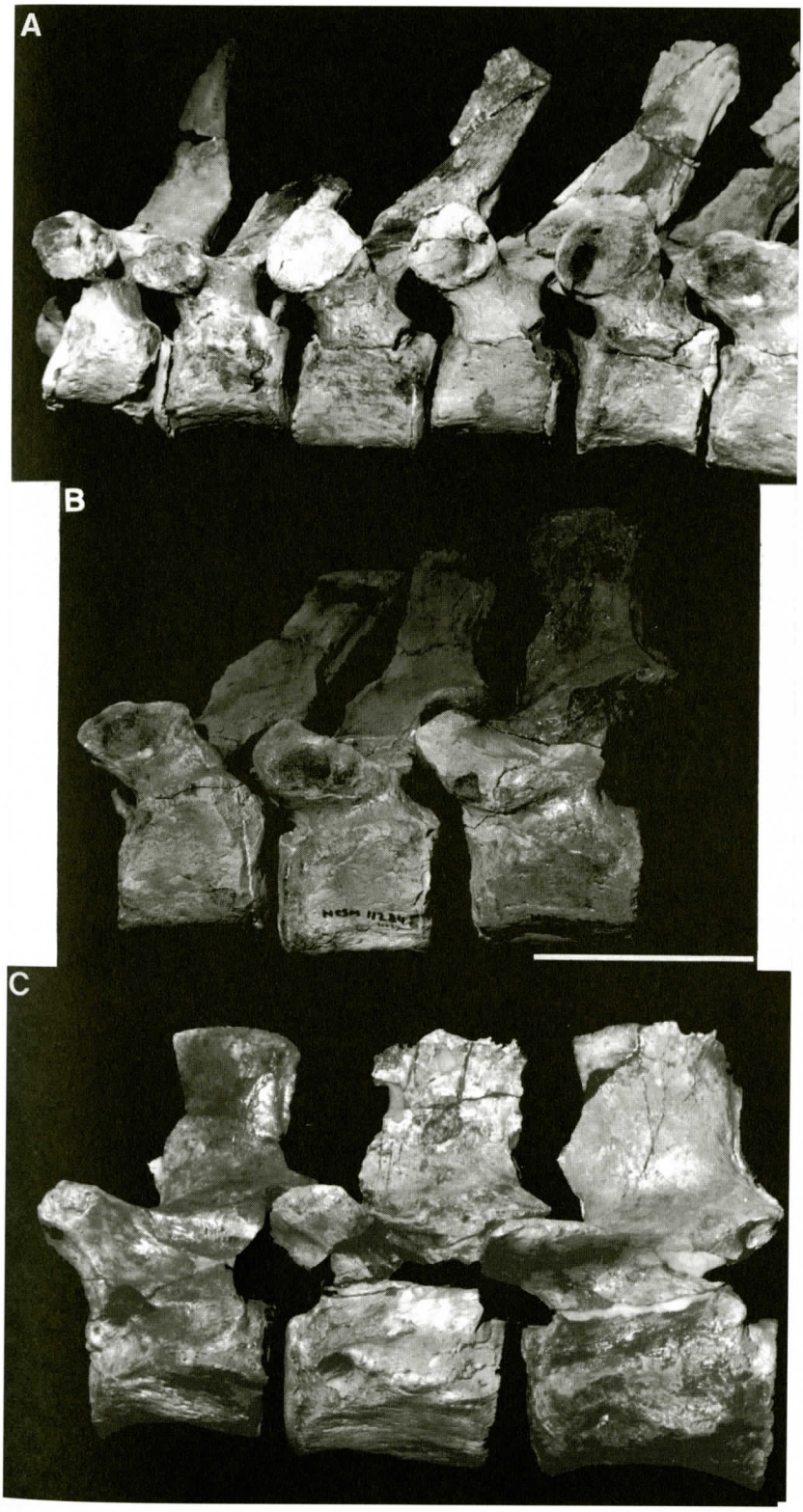

Figure 3. Thoracic vertebrae of *Eocetus wardii* in lateral view, NCSM 11284. A, T1 to T6; B, T7 to T9; C, T10 to T12. The neural spine of T5 and most of T6 are truncated on the right margin of A Scale bar is 10 centimeters.

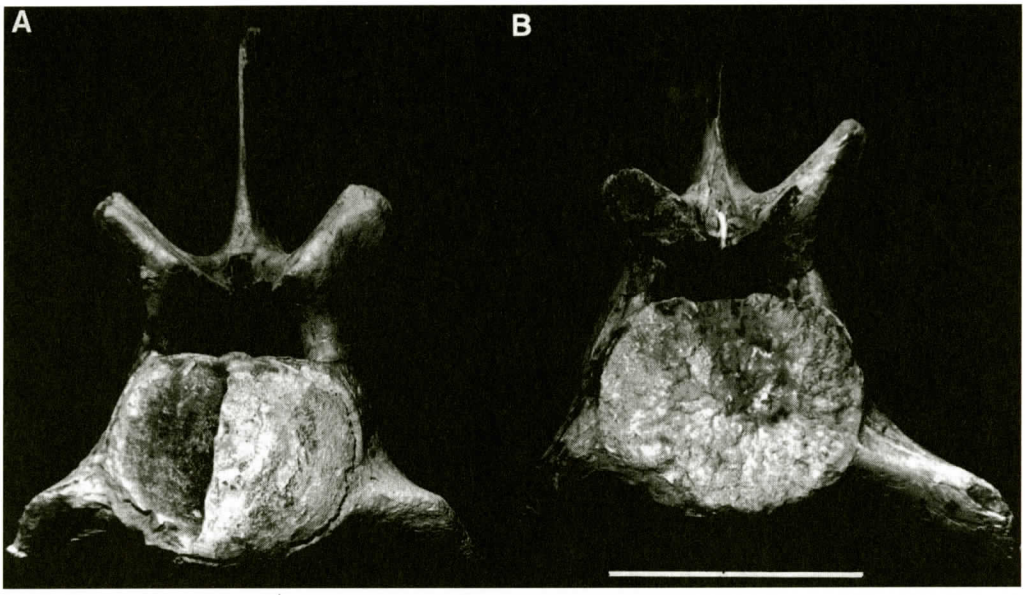

Figure 4. T10 and T12 of *Eocetus wardii* in anterior view, NCSM 11284. A, T10; B, T12. Scale bar is 10 centimeters.

(Harris and Zullo, 1991). It is not clear whether all of the newly referred specimens of *Eocetus wardii* listed above are from the Comfort Member or not, but they are all from quarries where the Comfort Member is exposed.

Description. — New specimens listed above have been identified as *Eocetus wardii* based on the distinctive texture of the bone, the size and shape of the vertebral bodies, and the distinctively broad transverse processes and neural spines of the vertebrae (particularly the posterior thoracic and lumbar vertebrae).

Left and right stylohyals are known from NCSM 11284, but both are missing their distal ends. They are long and slender with concave, irregularly oval-shaped proximal ends. They are very similar in size and morphology to those of *Dorudon* (Uhen, 1996).

Previously, only a few vertebrae from the thoracic, lumbar, and caudal regions were known from *Eocetus* (Uhen, 1999). Specimen NCSM 11284 includes a complete series of vertebrae from the seventh cervical (C7) to the second lumbar (L2), including twelve thoracic vertebrae between C7 and L1. Vertebral measurements are listed in Table 1. The specimen also includes a single posterior caudal vertebra. The seventh cervical vertebra (Figure 1) has broadly expanded transverse processes that lack foramina. The neural spine is short relative to the thoracic vertebrae.

The first thoracic vertebra has an oval anterior epiphysis that is similar in shape to the posterior epiphysis of C7. The unusual bone texture of *Eocetus* makes it difficult to distinguish costal articular surfaces on the vertebrae, but it appears that T1 to T7 have two articular surfaces on each vertebra (Figures 2 and 3), and T8 to T12 (Figures 3 and 4) have one articular surface on each vertebra. The thoracic vertebrae increase in length from anterior to posterior and are distinctly elongate by T10 (Figure 3). T8 is the anticlinal vertebra. The anterior-posterior length of the neural spines increases in proportion to the length of the vertebral bodies.

The lumbar vertebrae of NCSM 11284 have anteroposteriorly long transverse processes and neural spines, as well as large prezygapophyses. The transverse processes of L1 lack a distinct notch between the anterior margin of the processes themselves and the body of the vertebra. Both L1 and L2 have short transverse processes when compared to the lumbar vertebra described from the type specimen of *Eocetus wardii*, USNM 310633. This suggests that transverse process length increases from anteri-

Figure 5. L1 and L2 of *Eocetus wardii*, in lateral view, NCSM 11284. Scale bar is 10 centimeters.

or to posterior, which is also supported by the fact that L2 has longer transverse processes than L1 (Figure 5).

The caudal vertebra from NCSM 11284 is small (around 106 mm long, missing both epiphyses) when compared to the caudal vertebra described from the type specimen of Eocetus wardii, USNM 310633, probably because the USNM 31063 caudal is from the anterior caudal region, and the NCSM 11284 caudal is from the posterior caudal region. A pair of chevron articular facets is present on the ventral side of the body of the vertebra on the anterior end (Figure 6). Neither transverse process is perforated by a foramen, and the vertebra is not dorsoventrally compressed This may indicate that a tail fluke was not present in *E. wardii* since dorsoventrally compressed posterior caudal vertebrate are indicative of a tail fluke. It is also possible that the vertebra was simply anterior to the dorsoventrally compressed vertebrae of the tail stock. Without a more complete caudal vertebral series, it is impossible to know with certainty.

Two sternal elements of *Eocetus* are known from NCSM 11284, the manubrium and a mesosternal element. The manubrium is similar to that of *Georgiacetus vogtlensis*, broadly "T" shaped, with a broad anterior end, a narrow anteroposteriorly oriented body, and a slightly flared posterior end (Figure 7 A and B). The lateral margins of the anterior and posterior ends have rib articular surfaces. The body is dorsoventrally thickened at about the anteroposterior midline. The mesosternal element is unlike mesosternal elements known from other archaeocetes. It is generally rectangular in shape, elongate anteroposteriorly (Figure 7 A and B). It is dorsoventrally flat, but thickened along its margins. Both ends of the mesosternal element

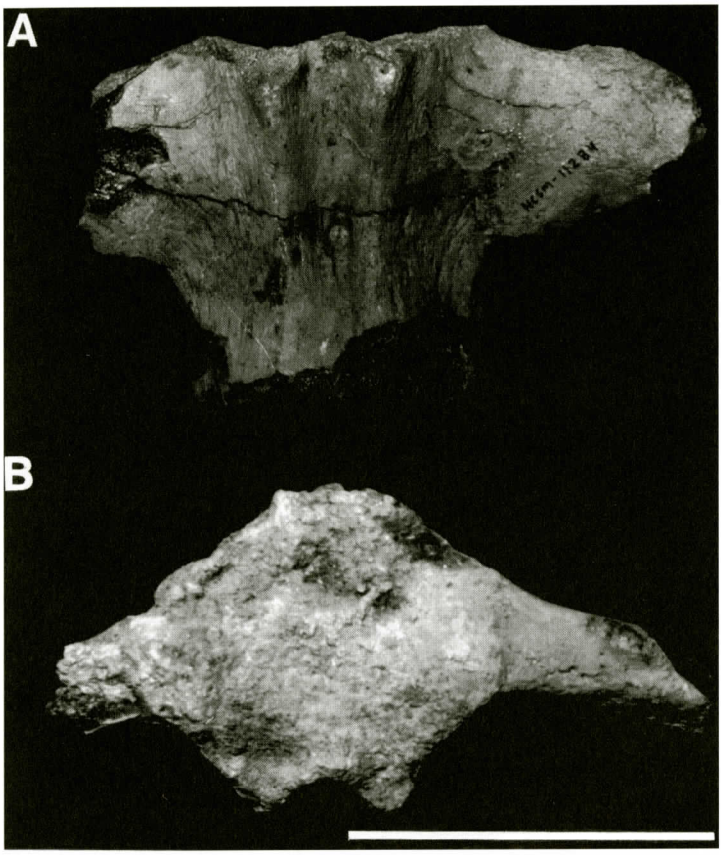

Figure 6. Caudal vertebra of *Eocetus wardii*, NCSM 11284. A, ventral view; B, anterior view. Scale bar is 10 centimeters.

have a rough, spongy surface texture indicative of a cartilaginous articular surface. On the lateral margins, about one third of the length from one end is a pair of rib articular surfaces. This end is presumed to be the anterior end, but it is not entirely clear that this is the case. The lateral corners of the presumed posterior margin also bear rib articular surfaces. It is unclear how this element relates to the manubrium.

Seven ribs of *Eocetus wardii*, specimen NCSM 11284, are relatively intact (Figure 7C). One of these ribs is short, stout, and has both a capitulum and tuberculum, indicating that it is probably the first rib. Others are considerably longer and bear both a capitulum and tuberculum, while other long ribs have only the capitulum. From the articular surfaces on the complete thoracic vertebral series, it appears that there are 12 pairs of ribs, and that the first seven ribs have both capitula and tubercula, while the remaining 5 have only capitula. The ribs of NCSM 11284 have the same layered cortical structure described in the type specimen of *E. wardii* (Uhen, 1999).

The scapula of *Eocetus* is unlike those of terrestrial mesonychians and is similar to those of basilosaurids and later cetaceans. It is fan shaped with a prominent spine (Figure 8). Both the acromion and coracoid processes and the glenoid fossa are missing from the specimen. It has a large supraspinous fossa and a greatly expanded infraspinous fossa when compared to the scapulae of mesonychians. What is preserved is very similar in size and shape to the scapula of *Dorudon* and other basilosaurids (see Uhen, 1998b).

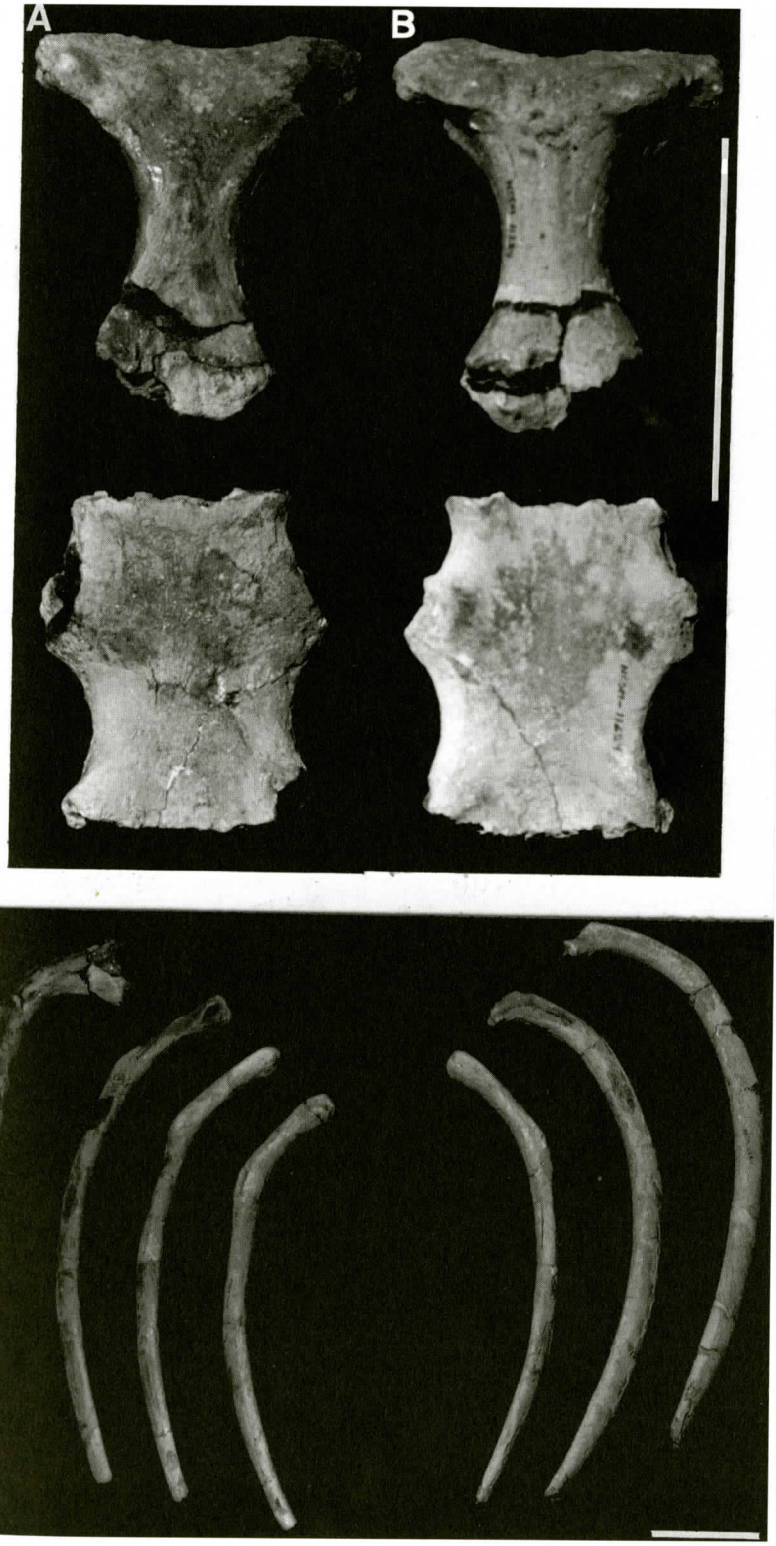

Figure 7. Sternum and ribs of *Eocetus wardii*, NCSM 11284. A, manubrium and mesosternal element in ventral view; B, manubrium and mesosternal element in dorsal view; C, ribs. Scale bar is 10 centimeters.

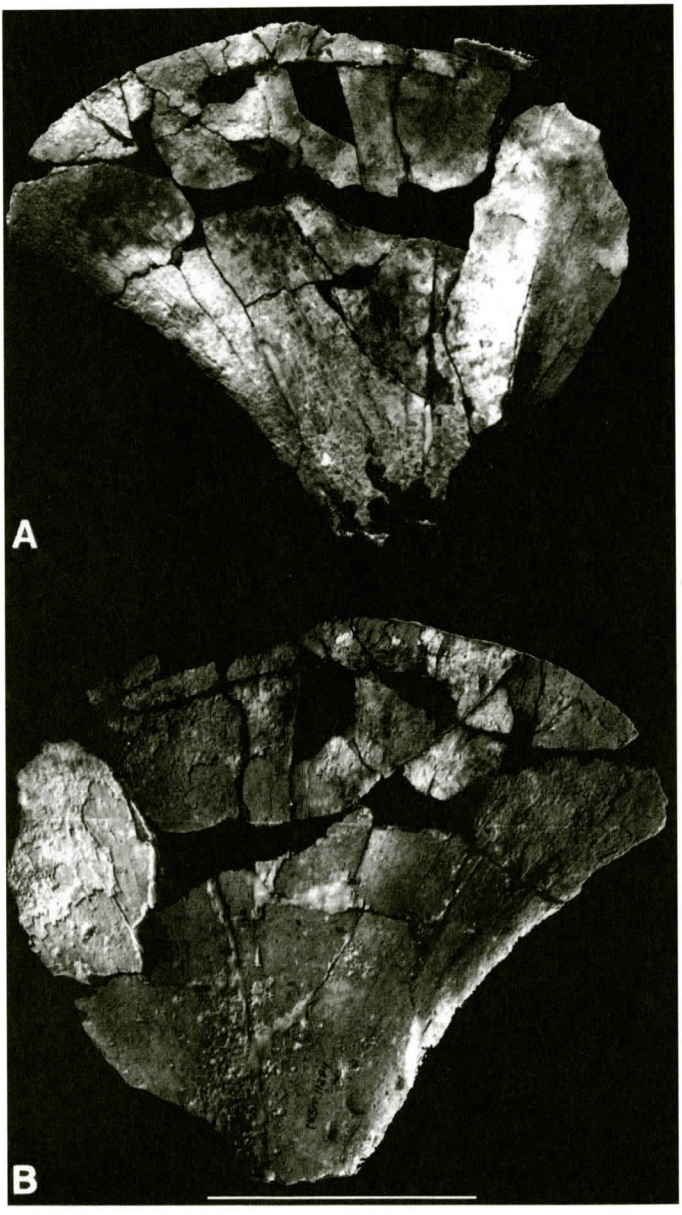

Figure 8. Right scapula of *Eocetus wardii*, NCSM 11284. A, lateral view; B, medial view. Scale bar is 10 centimeters.

DISCUSSION

Eocetus appears to combine some features common to protocetids with some features found otherwise only in basilosaurids. The scapula of *Eocetus* is similar in size and shape to that of *Dorudon atrox* (Uhen, 1998b). Scapulae of other protocetids (from as yet unpublished specimens) are much more narrow and lack expanded infraspinous fossae. The innominate of *Eocetus* is somewhat like those of basilosaurids in that the acetabulum is small and irregularly formed. The acetabulum of the innominate of *Eocetus* is not as small as in *Basilosaurus* (either absolutely or relatively) and some specimens of *Basilosaurus* and the newly described *Chrysocetus* as well have well-formed acetabula (Gingerich and others, 1990; Uhen and Gingerich, 2001). In addition, the narial fossa of *Eocetus* is relatively far back on the skull (above P^1 or slightly posterior to it, a feature shared with *Georgiacetus* as well) when compared to other protocetids.

What is known of the teeth of *Eocetus* suggests that they are similar to those of other protocetids. An upper molar attributed to the same stratigraphic horizon as the type specimen of *Eocetus schweinfurthi* (Gingerich, 1992) retains three roots and a worn protocone, both features found only in protocetids. Unfortunately, this molar is now missing. The anterior teeth (incisors plus canines) are more robust than the anterior teeth of similarly sized basilosaurids with better developed ornamentation. The cheek teeth lack multiple accessory denticles that are present in basilosaurid cheek teeth.

Elongation of the trunk vertebrae of *Eocetus* has led to speculation that *Eocetus* may share a special relationship with *Basilosaurus*, which also has elongate trunk vertebrae (Kellogg, 1936). Many features of *Eocetus* suggest that this is

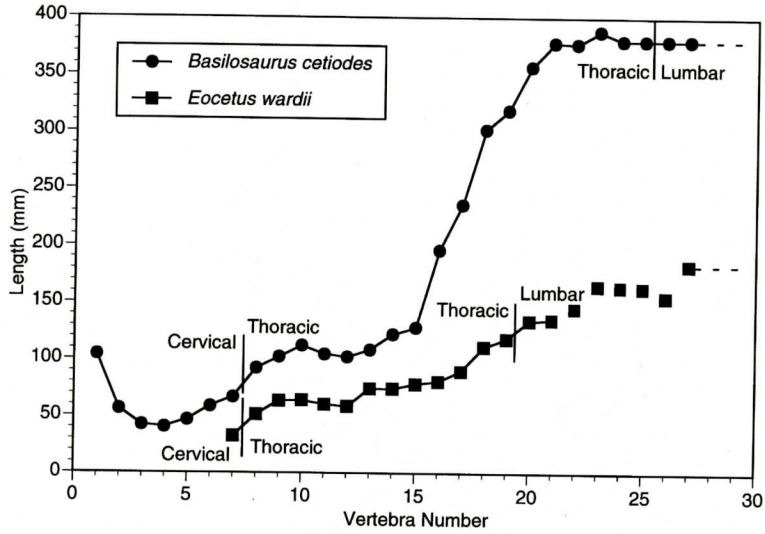

Figure 9. Plot of vertebral lengths of *Basilosaurus cetoides* and *Eocetus wardii*. Note that the vertebrae of *Basilosaurus* are larger and much more elongate in the posterior thoracic and lumbar region than those of *Eocetus*. The last six lumbar vertebrae shown for *Eocetus* are not in sequence with the others.

unlikely. First, the degree of elongation seen in *Eocetus* does not compare with that seen in *Basilosaurus*. Figure 9 shows that while the vertebrae of *Basilosaurus* increase in length dramatically in the mid-thorax, the vertebrae of *Eocetus* increase in length only gradually from the mid-thorax through the anterior lumbar region, and not to nearly the same extent as in *Basilosaurus*. Second, *Basilosaurus* does not have the same distinctive bone texture and layered bone growth as that seen in *Eocetus* (Uhen, 1999). Lastly, the style of vertebral elongation seen in *Basilosaurus* is different from that seen in *Eocetus*. The transverse processes and neural spines of *Basilosaurus* are not elongate in proportion to the length of the vertebral bodies as they are in *Eocetus* (Figure 5). These features, along with the large number of synapomorphies that *Basilosaurus* shares with other basilosaurids, indicate that *Basilosaurus* and *Eocetus* have no particularly close affinity.

PHYLOGENETIC RELATIONSHIPS OF *EOCETUS*

Eocetus falls within the paraphyletic archaeocete family Protocetidae as delimited by Thewissen and others (1996). The skull of *Eocetus* also shows more derived characters than members of the families Pakicetidae and Ambulocetidae including a frontal shield that has a straight posterior edge. *Eocetus* lacks the derived characters found in the family Remingtonocetidae. *Eocetus* does not belong in the family Basilosauridae, because it lacks many of the synapomorphies of that group including multiple, large accessory denticles on the cheek teeth, loss of M^3, and small, reoriented innominate.

A stratocladistic analysis that included 70 morphological characters and a single stratigraphic character was performed on 16 cetacean taxa, which included hypothetical ancestral mysticetes and odontocetes, and three additional mesonychine outgroup taxa (following Uhen, 1999). Stratocladistics uses morphologic data in the same way that conventional cladistic analysis does, but it also allows the relative stratigraphic occurrence of fossil taxa to be included in the analysis. This method treats *ad hoc* hypotheses that explain away missing fossils in the same way as *ad hoc* hypotheses of homoplasy. Hypotheses of relationships that minimize overall appeals to *ad hoc* hypotheses

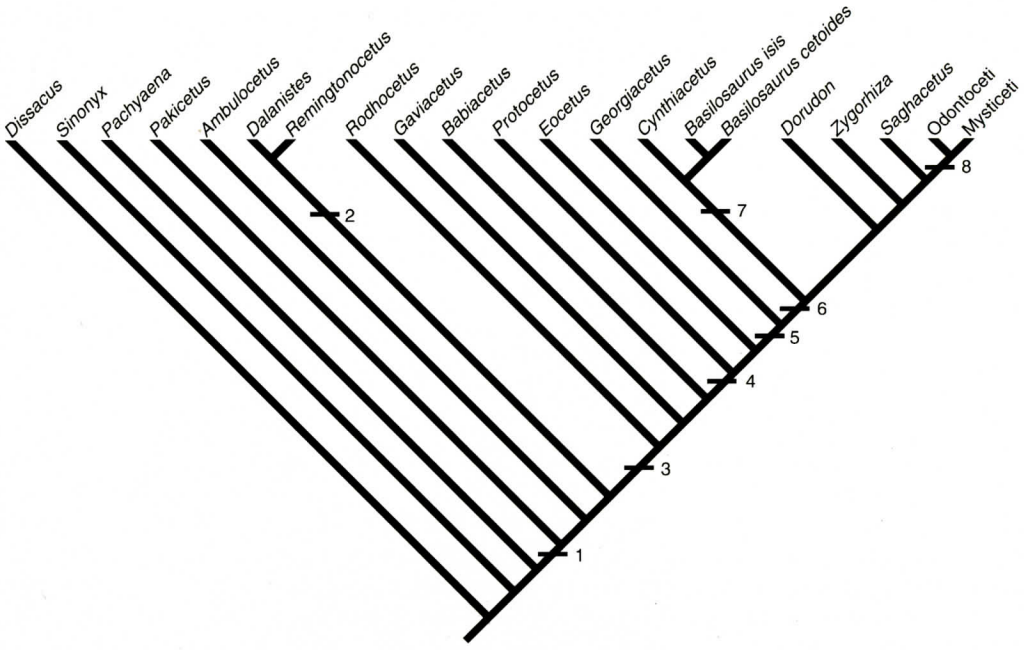

Figure 10. Cladogram of selected archaeocete taxa (after Uhen, 1999). This cladogram shows that *Eocetus* and *Georgiacetus* are more closely related to basilosaurid archaeocetes than any other protocetines. Many of the characters that make *Georgiacetus* group closely to the basilosaurids (incipient accessory denticles on the cheek teeth, reduced M^3, lack of a bony attachment of the innominate to the vertebral column) are unclear in *Eocetus*. Notable character state transitions are marked on the cladogram are as follows: 1—presence of a pachyosteosclerotic bulla; 2—elongation of the skull and fused mandibular symphysis; 3—reduction of the hind limb, presence of unfused sacral vertebrae; 4—single sacral vertebra; 5—innominate not attached to the vertebral column; 6—very reduced hind limb; loss of M^3, accessory denticles on cheek teeth well-developed; 7—very large body size; 8—monophyodonty.

are preferred by stratocladistics (Fisher, 1992). Stratocladistics also allows some taxa to be evaluated as ancestors, but ancestor-descendant relationships were not explored in this analysis.

All taxa were analyzed at the generic level in this analysis. The data matrix and character descriptions follow those listed in Uhen (1999) with the following additional data for *Eocetus wardii*: Character 38, number of thoracic vertebrae 1 (10-15); Character 47, posterior caudal vertebrae elongate, 1 (yes); Character 48, infraspinous fossa on scapula, 2 (large); Character 49, coracoid process oriented anteriorly, 1 (yes); Character 50, acromion process oriented anteriorly, 1 (yes). The initial conventional cladistic analysis was performed using the heuristic search option in PAUP 4.0b3 (Swofford, 2000). Search trees were built using a random addition sequence and 1000 repetitions of the search were performed.

Excluding the outgroup taxa, the result of the conventional cladistic analysis using only the morphologic data was nine morphologically most parsimonious trees of length 140. The nine trees in the set had consistency indices of 0.754 and retention indices of 0.817. After performing the conventional cladistic analysis a stratigraphic character was added in MacClade 3.0.7 (Maddison and Maddison, 1992). The result after the inclusion of stratigraphy was a single overall most parsimonious cladogram that is shown in Figure 10, which has an identical topology to the one reported by Uhen (1999). The position of *Eocetus* is well within a group of protocetines (*Gaviacetus*, *Babiacetus*, *Protocetus*, *Georgiacetus*). *Eocetus* and *Georgiacetus*

are the protocetids that are most closely related to the basilosaurids. Derived characters seen in *Georgiacetus* but unclear in *Eocetus* are: incipient accessory denticles on the cheek teeth and a reduced M^3. *Eocetus* is unlikely to be ancestral to the Basilosauridae due to the highly autapomorphic histology of the bone and shape of the vertebral processes.

SUMMARY AND CONCLUSIONS

New material of the protocetid archaeocete *Eocetus wardii* from North Carolina shows that *Eocetus* shares some derived characteristics of basilosaurids, but that is highly autapomorphic. This new material includes: a series of vertebrae from C7 to L2, including twelve thoracic vertebrae; a distal caudal vertebra; ribs and rib fragments; a manubrium and mesosternal element, and a nearly complete right scapular blade.

The new series of vertebrae clearly shows that while *Eocetus* has elongate trunk vertebrae that are similar to those of *Basilosaurus*, they are not nearly as elongate and that the pattern of vertebral elongation differs in these two taxa. The distal caudal vertebra is not dorsoventrally compressed, which may suggest that *Eocetus* lacked a tail fluke. The manubrium of *Eocetus* is similar to that of *Georgiacetus*, but the mesosternal element is unlike those of any other archaeocete. The scapular blade of *Eocetus* is similar to those of basilosaurid archaeocetes and unlike those known from other protocetids.

One of the most puzzling aspects of *Eocetus wardii* is its geographic distribution. *E. wardii* is known only from the Castle Hayne Formation in North Carolina. Despite intense searching in other nearby areas (South Carolina, Georgia, Alabama, Mississippi), no specimens of *E. wardii* have been discovered outside of North Carolina. This despite the fact that even individual bones and bone fragments of *E. wardii* are easily identified based upon the distinctive bone texture. Perhaps *E. wardii* had very narrow environmental preferences that were only present in North Carolina during the late middle Eocene. This idea could be tested by attempting to place all of the specimens of *E. wardii* in a more detailed stratigraphic framework to see if the environments in which they were deposited show any significant differences from other environments in which North American protocetids are recovered.

ACKNOWLEDGMENTS

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REFERENCES CITED

Albright, L. B. III, 1996, A protocetid cetacean from the Eocene of South Carolina: Journal of Paleontology, v. 70, p. 519-522.

Andrews, C. W., 1920, A description of new species of zeuglodont and of leathery turtle from the Eocene of southern Nigeria: Proceedings of the Zoological Society of London, v. 1919, p. 309-319.

Brisson, A. D., 1762, Regnum Animale in Classes IX distributum sive synopsis methodica. Edito altero auctior: Theodorum Haak, p. 294.

Feldmann, R. M., Bice, K. L., Hopkins, C. S., Salva, E. W., and Pickford, K., 1998, Decapod crustaceans from the Eocene Castle Hayne Limestone, North Carolina: Paleoceanographic implications: Journal of Paleontology, The Paleontological Society Memoir, v. 48, p. 1-28.

Fisher, D. C., 1992, Stratigraphic Parsimony, *in* Maddison, W. P. and Maddison, D. R., MacClade 3.0 Manual: Sinauer Associates, p. 124-129.

Flower, W. H., 1883, On the arrangement of the orders and families of existing Mammalia: Proceedings of the Zoological Society of London, v. 1883, p. 178-186.

Fraas, E., 1904, Neue Zeuglodonten aus dem unteren Mittelecocän vom Mokattam bei Cairo: Geologische und Palæontologische Abhandlungen, Jena, Neue Folge, v. 6, p. 199-220.

Geisler, J., Sanders, A. E., and Luo, Z., 1996, A new protocetid cetacean from the Eocene of South Carolina, U.S.A; phylogenetic and biogeographic implications: Sixth North American Paleontological Convention Abstracts of Papers. The Paleontological Society Special Publication, v. 8, p. 139.

Gingerich, P. D., 1992, Marine mammals (Cetacea and Sire-

nia) from the Eocene of Gebel Mokattam and Fayum, Egypt: stratigraphy, age and paleoenvironments: University of Michigan Museum of Paleontology Papers on Paleontology, v. 30, p. 1-84.

Gingerich, P. D., Arif, M., and Clyde, W. C., 1995, New archaeocetes (Mammalia, Cetacea) from the middle Eocene Domanda Formation of the Sulaiman Range, Punjab (Pakistan): Contributions from the Museum of Paleontology, University of Michigan, v. 29, p. 291-330.

Gingerich, P. D., Raza, S. M., Arif, M., Anwar, M., and Zhou, X., 1993, Partial skeletons of *Indocetus ramani* (Mammalia, Cetacea) from the lower middle Eocene Domanda Shale in the Sulaiman range of Punjab (Pakistan): Contributions from the Museum of Paleontology, University of Michigan, v. 28, p. 393-416.

Gingerich, P. D., Smith, B. H., and Simons, E. L., 1990, Hind limbs of Eocene *Basilosaurus*: Evidence of feet in whales: Science, v. 229, p. 154-157.

Halstead, L. B. and Middleton, J. A., 1974, New material of the archaeocete whale, *Pappocetus lugardi* Andrews, from the middle Eocene of Nigeria: Journal of Mining and Geology, v. 8, p. 81-85.

Harris, W. B. and Laws, R. A., 1997, Paleogene stratigraphy and sea-level history of the North Carolina Coastal Plain: Global coastal onlap and tectonics: Sedimentary Geology, v. 108, p. 91-120.

Harris, W. B. and Zullo, V. A., 1991, Eocene and Oligocene stratigraphy of the outer Coastal Plain, *in* Horton, J. W. Jr. and Zullo, V. A., eds., The Geology of the Carolinas: The University of Tennessee Press, p. 251-262.

Hazel, J. E., Bybell, L. M., Edwards, L. E., Jones, G. D., and Ward, L. W., 1984, Age of the Comfort Member of the Castle Hayne Formation, North Carolina: Geological Society of America Bulletin, v. 95, p. 1040-1044.

Hulbert, R. C., Petkewich, R. M., Bishop, G. A., Bukry, D., and Aleshire, D. P., 1998, A new middle Eocene protocetid whale (Mammalia: Cetacea: Archaeoceti) and associated biota from Georgia: Journal of Paleontology, v. 72, p. 907-927.

Kellogg, R., 1936, A Review of the Archaeoceti: Carnegie Institute of Washington Special Publication, v. 482, p. 1-366.

Maddison, W. P. and Maddison, D. R., 1992, MacClade: Analysis of Phylogeny and Character Evolution Version 3.07: Sinauer Associates, Inc., p. 398.

McLeod, S. A. and Barnes, L. G., 1996, The systematic position of *Pappocetus lugardi* and a new taxon from North America (Archaeoceti: Protocetidae): Sixth North American Paleontological Convention Abstracts of Papers. The Paleontological Society Special Publication, v. 8, p. 270.

Sahni, A. and Mishra, V. P., 1975, Lower Tertiary vertebrates from western India: Monograph of the Palaeontological Society of India, v. 3, p. 1-48.

Stromer, E., 1908, Die Archaeoceti des Ägyptischen Eozäns: Beiträge zur Paläontolgie und Geologie Österreich-Ungarns und des Orients, v. 21, p. 1-70.

Swofford, D. L., 2000, PAUP* Phylogenetic Analysis Using parsimony (* and Other Methods): Sinauer Associates, p. Version 4.0b3.

Thewissen, J. G. M., Madar, S. I., and Hussain, S. T., 1996, *Ambulocetus natans,* an Eocene cetacean (Mammalia) from Pakistan: Courier Forschungsinstitut Senckenberg, v. 191, p. 1-86.

Trivedy, A. N. and Satsangi, P. P., 1984, A new archaeocete (whale) from the Eocene of India: Abstracts of the 27th International Geological Congress, Moscow, v. 1, p. 322-323.

Uhen, M. D.1996, *Dorudon atrox* (Mammalia, Cetacea): Form, Function, and Phylogenetic Relationships of an Archaeocete from the Late Middle Eocene of Egypt [Ph.D. dissertation]: Ann Arbor, University of Michigan, 608 p.

Uhen, M. D., 1998a, New protocetid (Mammalia, Cetacea) from the late middle Eocene Cook Mountain Formation of Louisiana: Journal of Vertebrate Paleontology, v. 18, p. 664-668.

Uhen, M. D., 1998b, Middle to Late Eocene Basilosaurines and Dorudontines, *in* Thewissen, J. G. M., ed., The Emergence of Whales: Plenum Press, p. 29-61.

Uhen, M. D., 1999, New species of protocetid archaeocete whale, *Eocetus wardii* (Mammalia, Cetacea), from the middle Eocene of North Carolina: Journal of Paleontology, v. 73, p. 512-528.

Uhen, M. D. and Gingerich, P. D., 2001, New genus of dorudontine archaeocete (Cetacea) from the middle-to-late Eocene of South Carolina: Marine Mammal Science, v. 17, p. 1-34.